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GEOLOGY OF THE CARDIGAN QUADRANGLE, NEW HAMPSHIRE

BY KATHARINE FOWLER-LUNN AND LOUISE KINGSLEY

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INTRODUCTION

LOCATION

The Cardigan quadrangle lies in the west-central part of New Hampshire, between latitude 43°30′ and 43°45′ north and longitude 71°45′ and 72°00′ west. Newfound Lake, at an altitude of 586 feet, is the only large lake in the district. The higher sections are dominated by Mt. Cardigan, which rises to an altitude of 3121 feet. Although this mountain and a few neighboring summits are barren, the higher sections in most of the area are heavily wooded. Danbury, Alexandria, Hebron, Orange, and Grafton are small farming centers within the quadrangle. Three large towns lie just beyond the boundaries—Plymouth, three miles to the northeast; Bristol, one mile southeast; and Canaan, one mile west.

STATEMENT OF PROBLEM

The geology of the Cardigan quadrangle has been studied with special reference to the problems associated with the intrusion of the Kinsman quartz monzonite or porphyritic gneiss. The results of the structural and petrographic study of the rocks of the region are presented, in the following pages, together with a brief summary of the geologic history.

The Cardigan region was chosen as a field for study because of the problems offered by the porphyritic gneiss. It was decided to map the whole quadrangle, rather than concentrate merely on the Kinsman gneiss, so that the results would fit in with other quadrangles in New Hampshire, which were being mapped at the same time, or which might be mapped in the future. This precedent of mapping a quadrangle as a unit had been established by the geologists already working farther north, with the hope that, in the absence of a geological survey in New Hampshire, the geology of the State might eventually be brought up to date by independent investigators.

ACKNOWLEDGMENTS

The writers wish to express their appreciation of the co-operation of Dr. Marland Billings, of Harvard University, who made visits on various occasions with groups working in neighboring quadrangles. Type localities were visited so that an agreement could be reached as to nomenclature and correlation.

FIGURE 1.—*Location of Cardigan (New Hampshire) quadrangle*

The work of Dr. Billings and his associates in various parts of New Hampshire has been invaluable as an aid in the interpretation of the history of the area.

Mr. Carleton Chapman, of Harvard University, who has completed the map of the Mascoma quadrangle to the west; Mr. Jarvis Hadley, of Tufts College, who has completed the map of the Mt. Cube quadrangle to the northwest; and Mr. Lincoln Page, of the University of Minnesota, who is working in the Rumney quadrangle to the north, were consulted and were found to be in agreement with the field mapping.

Valuable suggestions about the structure were given by Dr. Robert Balk, of Mount Holyoke College, who spent two days in the field during the preliminary studies.

The writers are indebted to Dr. Billings and the members of the geology department of Wellesley College for reading the manuscript.

PREVIOUS WORK IN THE CARDIGAN REGION

Since the publication of Hitchcock's atlas of New Hampshire in 1878, little geological work has been done in the Cardigan region aside from reports dealing with the mica and garnet mines associated with the numerous pegmatites, and a paper by Daly in 1897 on the porphyritic gneiss of New Hampshire.

FIELD WORK

The entire summers of 1935 and 1936 were spent in the field. Contacts were traced where possible, or were plotted from outcrop to outcrop across the region. The nature of the intrusions made detailed pace and compass traverse unnecessary in most cases. About eighty petrographic slides were studied from specimens collected in the field.

PETROLOGY AND ROCK DESCRIPTION

GENERAL LITHOLOGIC FEATURES

Almost all the rocks of the Cardigan quadrangle belong to the New Hampshire magma series and the Littleton schists of Devonian age, which they intrude. In the northwest corner, a small belt of Ammonoosuc volcanics of probable Ordovician age, is cut by a granite of the Oliverian magma series. The age sequence is discussed fully by Billings in his paper on the Paleozoic age of the rocks of central New Hampshire. The description of the rock types, which follows, will start with the oldest of the groups studied in the region.

AMMONOOSUC VOLCANICS

The Ammonoosuc volcanics, which are intruded by a member of the Oliverian magma series, named the Smarts Mountain granite by Hadley, are older than the Littleton schists and the New Hampshire magma series.

Billings considers them to be of Ordovician age. They consist of amphibolites alternating with highly metamorphosed siliceous volcanic rocks, all of which have been injected by granites of the Oliverian magma series, as well as by basic dikes of a much later age.

LITTLETON SCHISTS

The Littleton schists are continuous from the Moosilauke quadrangle, through the Rumney quadrangle, into the Cardigan area. They lie on both sides of the Kinsman intrusive mass, making two long belts running from north to south. The whole region is in the high-grade zone, as the schists are highly metamorphosed and contain sillimanite. Middle-grade and low-grade rocks are found farther west, toward the Connecticut River. Billings' studies in the Moosilauke quadrangle have proved the schists to be Devonian in age, thereby correlating them with the Littleton schists farther west. It is for this reason that the name "Littleton schist" has been adopted for the Cardigan region, rather than "Montalban group" or "mica schists with fibrolite", as used by Hitchcock in his atlas.

Within this quadrangle, the Littleton formation consists of schists and quartzites. The schists everywhere are rich in sillimanite. The quartzose members vary from fine- or medium-grained quartzites and garnetiferous quartz-mica schists to rocks that appear much like fine-grained granites. Among the schists, all gradations may be found from rocks in which only quartz and mica show in the hand specimen, through garnet-sillimanite schists.

The large proportion of schist containing sillimanite crystals at least an inch long is the most distinctive feature of the Littleton formation. Wherever the crystals are numerous but too small to be seen clearly, they give the rock a silvery luster and fibrous, splintery texture. The size of the crystals is related to the original sedimentary layers. Within a single bed, they remain constant in size, yet they vary conspicuously from bed to bed. There is practically no area of any size where the argillaceous layers are free from them. In general, the sillimanite crystals are largest near the major intrusions or where lit-par-lit injection is extensive. They are particularly well developed in a mile-wide belt directly west of, and parallel to, the main Kinsman intrusion.

Microscopic examination shows that the arenaceous layers of the Littleton formation contain quartz, extremely varying amounts of orthoclase and andesine, biotite that has locally altered to chlorite, and a little muscovite. Small garnets are characteristic. Brown tourmaline, as well as magnetite, pyrite, apatite, and zircon, are accessories. Quartz grains have mosaic and sutured textures. Micas show an almost regular parallelism, though bands of more closely spaced flakes still record original varia-

tion in the composition of the sedimentary layers. Garnets are small and hypidiomorphic. Many of them are aligned in the foliation plane.

In the argillaceous layers, biotite is the dominant mica, although muscovite is found as shreds in some of the shear planes. Quartz and varying amounts of orthoclase form narrow lenses. Garnets are common, are generally granular, and have altered in part to biotite or chlorite. Sillimanite is not found everywhere; some exposures show swarms of fine needles in conspicuous flow lines or as large crystals. It is associated with muscovite. Finely divided and contorted muscovite in shear planes is, in some places, crowded with minute sillimanite needles, along the flow lines. Many of these needles penetrate quartz, extending uninterruptedly across the quartz-mica contacts; they have not been granulated. Other sections, in which cores of large crystals are fractured and surrounded by needles comparable in size to the fractured fragments in whole crystals, suggest that granulation has strewn the fragments through the schists.

The association of sillimanite with muscovite, and the universal presence of orthoclase, lead to the conclusion that at least part of the sillimanite, probably the needle-like swarms, have been derived from the breakdown of the mica. "Knots" of muscovite, quartz, and orthoclase, resembling sillimanite crystals in form, suggest that some of the sillimanite has undergone late hydrothermal alteration. Radiating clusters of chlorite occupy circular areas, which, from their outline and the position of the remaining fragments of garnet, must have been augen.

Although garnets are common throughout most of the Littleton formation, the richest garnet deposits are limited to a narrow belt in the south-central part of the quadrangle. Here, where they have been mined in two quarries, one of which is 1½ miles west, and the other 4 miles southwest of Danbury, they make up as much as fifty per cent of the rock. These garnet deposits have been discussed by Louis Conant. The garnets are found in long, narrow roof-pendants of schist, which have been intensely sheared and injected lit-par-lit by the invading Kinsman magma. They are augen, at the ends of which are found clear, unstrained quartz and orthoclase. Narrow bands of garnet-sillimanite-biotite schists alternate with bands of Kinsman quartz monzonite as well as with fine-grained granite. The origin of the individual layers is, consequently, difficult to determine in many places. The alternating bands vary from nearly microscopic size to several feet in width.

NEW HAMPSHIRE MAGMA SERIES

General description.—The New Hampshire magma series is made up of intrusions, which are probably of late Devonian age, although injection possibly continued into Carboniferous time. In the Cardigan region, three subdivisions have been recognized: the Bethlehem gneiss, the Kins-

man quartz monzonite, which is a porphyritic gneiss, and the Concord granite. The "Franconia" breccia is associated with the late stage of the Concord granite.

Bethlehem gneiss.—The Bethlehem gneiss crops out as a long, narrow belt in the northwestern part of the quadrangle, and can be traced continuously into the Mascoma, Mt. Cube, Rumney, Mt. Moosilauke, and Franconia quadrangles. Billings considers it to be one of the older members of the New Hampshire magma series. The name is taken from rocks described by Hitchcock near Bethlehem. In the Cardigan region, therefore, the term Winnipiseogee gneiss, as used in Hitchcock's atlas, is abandoned.

The rock has a medium grain, is gray, and is made up of quartz, white feldspar, biotite, and variable amounts of muscovite. It is the most strongly foliated igneous rock of the quadrangle. Although the foliation gives it an apparent coarse grain, most of the individual crystals are not more than one or two millimeters long. In certain areas, large crystals of feldspar, up to three centimeters in diameter, are common. Under such conditions, it closely resembles the Kinsman porphyritic quartz monzonite.

Microscopically, the quartz is seen either as interlocking grains with feldspar or as long, narrow lenses of pure quartz. The plagioclase is andesine and is almost free from zoning. Although many of the crystals show only plagioclase twinning, most of them are also twinned according to the pericline law. Potash feldspar is extremely variable in amount and character, showing either as small or as extremely large crystals of orthoclase or microcline. Both orthoclase and microcline may be seen in the same thin-section. Biotite crystals are subparallel and tend to be concentrated in shear zones. Some muscovite is intergrown with the biotite and resembles it in size and arrangement of crystals. It is found within the plagioclase and potash feldspars as exceptionally large flakes, distributed either evenly through the cores of the crystals or parallel to cleavage planes. The muscovite has an unusually small optic angle.

The accessory minerals are magnetite, pyrite, zircon, and apatite. In some sections, soda-lime feldspars have altered to sericite, and the potash feldspars remain unaltered. Some of the chlorite is an alteration product of biotite.

The texture of the rock gives clear evidence of movement during consolidation. Long, narrow lenticles of pure quartz or quartz and feldspar show varying degrees of strain, bending of twin lamellae, and granulation of crystal borders. The tendency for parallel biotite flakes to be concentrated between these lenticles emphasizes the effect of shearing. Lack of granulation of myrmekite suggests that movement ceased before its formation.

In general, the Bethlehem gneiss may be distinguished from the Kinsman quartz monzonite by a more uniform texture, finer grain, and more perfect foliation with less granulation.

Kinsman quartz monzonite.—The name, Kinsman quartz monzonite, is used for the large body of porphyritic gneiss that occupies the central part of the Cardigan quadrangle. On the Hitchcock map, it is called "porphyritic gneiss"; it was considered to be of sedimentary origin, representing sediments collected from "possibly the original crust of the earth."[1]

It was the oldest rock recognized in Hitchcock's time. In 1897, Daly described the gneiss as a porphyritic granite but did not give it a name.

In 1935, Conant used the term, Meredith granite, in this area, applying the term to the porphyritic gneisses over all the central part of New Hampshire, following the terminology of Billings in the North Conway quadrangle. But, as Billings has used the name Kinsman quartz monzonite in describing the porphyritic gneisses in Kinsman Notch, due north of the Cardigan region, the more restricted term, Kinsman quartz monzonite, is used in this report instead of Meredith granite. This distinguishes the belt doubtlessly continuous with the rocks of Kinsman Notch, to the north, from the other porphyritic gneisses east and west of it.

The quartz monzonite is coarsely foliated, medium-grained, and gneissoid in texture. It is composed of quartz, feldspar, biotite, and muscovite. Shearing and granulation have reduced the grain size so that few grains, except for the large orthoclase crystals, are more than one or two millimeters long. One of the most conspicuous and distinctive features of the rock is the abundance of these large potash feldspars. The largest are about five or six centimeters in length. Crystals of this size are common throughout the main body of the rock. Locally, they may be only one or two centimeters long. They show Carlsbad twinning, and are subhedral to anhedral in outline. Another conspicuous feature of the rock is the knots or stringers of light-colored constituents. These vary from short, thick, lens-shaped areas about two centimeters long, to long, thin, granulated stringers some of them ten centimeters in length where shearing has been most intense.

Microscopically, the composition of the Kinsman quartz monzonite is similar to that of the Bethlehem gneiss. It contains abundant quartz with andesine as the dominant, and in many specimens the only, feldspar of the finer-grained part of the rock. Potash feldspar, where present, is generally microcline, although some orthoclase has been noticed. In distribution, the potash feldspar is erratic, being found either as small

[1] C. H. Hitchcock: *Geology of New Hampshire*, vol. 1 (1874) p. 512.

FIGURE 1. SUMMIT

FIGURE 2. VIEW FROM BRIDGEWATER HILL
Looking west. Bear Mountain ridge in middle distance. Photo by Moody, Bristol, New Hampshire.

MT. CARDIGAN

FIGURE 1. KINSMAN QUARTZ MONZONITE
Showing parallelism of feldspar masses

FIGURE 2. POTHOLES ALONG THE COCKERMOUTH RIVER

FIGURE 3. LITTLETON SCHISTS ALONG THE FOWLER RIVER

OUTCROPS IN THE CARDIGAN QUADRANGLE

granulated particles or as large crystals with irregular outlines. Brown biotite and muscovite are found in about equal proportions, except near the contact with the schists, where biotite dominates. Muscovite, as in the Bethlehem gneiss, is found as platy intergrowths with biotite and as conspicuously large oriented crystals in soda-lime and potash feldspar, and in both forms is similar.

Accessory minerals are zircon, generally enclosed in biotite, apatite, magnetite, and pyrite. Large amounts of garnet and sillimanite are found near schist contacts where the magma has injected schists lit-par-lit.

The texture of the Kinsman quartz monzonite farthest from the sheared contacts and roof pendants is granoblastic. Nearer the schists, cataclastic textures dominate and are extremely well developed. Quartz forms long, narrow lenses, which are strained, fractured, and show granulated margins. Feldspars have broken more easily, forming a fine mosaic. Concentration of thin, parallel micas along the granulated shear zones has produced augen and granulated textures throughout the rock. The augen or quartz lenses between the schist lamellae are outlined by curving bands of sillimanite, biotite, and granulated garnets. Quartz introduced between the lamellae of the schist has been strained and shows a fine fracture pattern with interlocking or sutured margins.

Concord granite.—The name, Concord granite, has been used for the finer-grained, non-porphyritic, biotite-muscovite granites, for they resemble the granite in the quarries near Concord, New Hampshire. Hitchcock, in his atlas, also used this name for two patches of granite, which he showed on the west side of Newfound Lake, within the Cardigan quadrangle. This granite is the youngest of the New Hampshire magma series.

There is a large mass of Concord granite on the west side of Newfound Lake, as well as east and west of Tenney Mountain and in the vicinity of Jewell Hill. In the southwestern part of the quadrangle, south of Mt. Cardigan, large areas of the granite are found. There are also innumerable dikes of Concord granite cutting all the rocks of the region.

There are two phases of the Concord granite, an earlier, coarser type, and a late, fine-grained phase. Because they are closely related, it seems unnecessary to give them different names. They have been shown on the geological map as one unit.

The Concord granite is apparently a late differentiate of the same magma that produced the Kinsman and the Bethlehem intrusions. Its composition varies from an oligoclase-biotite granite to one in which the feldspar is largely microcline. Although the proportion of microcline in the coarser earlier intrusions of Concord granite is small in many

places, the finer-grained, late phases are everywhere rich in microcline.

The coarse phase of the rock is a gray, medium-grained, equi-granular granite, composed of sugary quartz, white feldspar, and biotite. There is distinct foliation, although it is much less conspicuous than in the older intrusions. The foliation increases noticeably near the contacts, and in some places is so intense that the rock resembles Bethlehem gneiss. The proportion of biotite and garnet also increases with proximity to the schists. At "The Ledges," on the west shore of Newfound Lake, where recent road construction has opened up large, fresh exposures, shadowy biotite-rich spots indicate partial assimilation of inclusions. Variable composition of the non-foliated granite, and the presence of indistinct xenoliths at "The Ledges," combined with unusual amounts of biotite near the contact with the schists, give evidence of partial solution of wall rock.

Microscopic examination shows that the composition of this coarse phase of the Concord granite is much like that of the Bethlehem and the Kinsman granites already described. The soda-lime feldspar is oligoclase rather than andesine. Microcline forms irregular patches, no larger than the biggest oligoclase crystals. The proportion of the two feldspars is extremely variable. Muscovite, as in the older granitic rocks, is exposed as large flakes intergrown with biotite or as crystals parallel to feldspar cleavages. The proportion and kind of accessories are likewise similar to those found in the older intrusions.

Eliminating the differences caused by large potash feldspar crystals, found only in the Kinsman quartz monzonite and the Bethlehem gneiss, the texture is similar to that of the older rocks. There is less evidence of strain, although some is recorded in the quartz and by bent feldspar lamellae. Only locally, or at the borders of the intrusion, is there appreciable granulation. A few fragmentary garnets and shreds of muscovite and sillimanite give evidence of solution of inclusions.

The fine-grained phases of the Concord granite include a group of binary granites, which inject the older rocks and show almost no foliation. In these, the grain size grades from about that of the earlier phase of the Concord granite to a distinctly fine-grained phase. In color, they are lighter gray than the older rocks. In a few places the feldspar is slightly pinkish.

The composition of the fine-grained phase is similar to the composition of the potash-rich coarser phase. Only rarely is oligoclase the dominant feldspar. Zones within the microcline show perthitic intergrowths of albite, which apparently had been in solid solution in the microcline. Conspicuous inclusions, mainly of plagioclase, are scattered through the microcline.

In texture, the fine-grained Concord granite is more nearly granitic than any of the older injections. Although quartz shows strain, there is no granulation, little, if any, bending of the micas, and almost complete absence of foliation, indicating its intrusion after deformation had ceased.

"Franconia" breccia.—A breccia, resembling the "Franconia" breccia of Hitchcock and the one described by Williams in the Franconia quadrangle, is exposed for several hundred feet along the road following the Fowler River, one mile west of the Crawford School, Alexandria, as well as in the river bed. It is the only outcrop of its kind within the quadrangle. It is an intrusive breccia, related to the contact of the schist and granite. Angular fragments of schist and non-porphyritic granite up to several feet in length are embedded in a matrix of the fine-grained late phase of the Concord granite. The development of large garnet aggregates in the schists of the brecciated zone is striking.

STRUCTURAL GEOLOGY

GENERAL STRUCTURAL FEATURES

The structural features of the rocks of the Cardigan quadrangle yield considerable information regarding the methods of injection of the igneous bodies and their effect upon the country rock. (See Plate 3 for details of structure and geologic boundaries, and cross-sections.)

The intrusion of the Kinsman quartz monzonite into the tightly folded Littleton schists, probably at the close of the Devonian period, is one of the key problems in the interpretation of the history of the area. A study of the structures of the Kinsman igneous mass shows that it has been injected in the form of an elongate sill, the source of the igneous material being to the southeast. The schists at the base of the sill are overturned to the west, as a result of the intense deformation accompanying the intrusion of the Kinsman.

The Bethlehem gneiss is likewise a sill-like mass, which has been injected into the Littleton schists and the Ammonoosuc volcanics, resting nearly horizontally on the latter.

The Concord granite, on the other hand, is not restricted to belts, as are the earlier intrusions. It has modified only slightly the previously developed structures.

The regional strike, N. 20°E., of the foliation planes in the Littleton schists and the Bethlehem and the Kinsman porphyritic gneisses is constant, maintaining this alignment through the Rumney quadrangle to the Moosilauke quadrangle.

AMMONOOSUC VOLCANICS

The Ammonoosuc volcanics make up the east limb of a low dome, the eastern edge of which is exposed south of Thompson Hill, in the north-

western part of the quadrangle. The bedding planes dip gently eastward, at an angle of about twenty degrees, the dip being parallel to the foliation of the overlying Bethlehem gneiss, which makes up Thompson Hill. The pattern of the contact plane at the base of Thompson Hill gives additional proof that at its western border the Bethlehem gneiss rests on the volcanics, at a low angle.

The Oliverian magma series, represented by a small amount of the Smarts Mountain granite exposed along the northwestern border of the quadrangle in the vicinity of Lary Pond, has intruded the Ammonoosuc series. It shows almost horizontal foliation and dips under the Ammonoosuc volcanics.

South of Lary Pond, the volcanic rocks and the granite are cut off by an east-west fault, which has displaced the southern section to the west. The presence of the fault is indicated by the abrupt termination of the amphibolites along the line of strike. Further proof of its existence is found in the Mascoma quadrangle.

LITTLETON SCHISTS

The Littleton schists are isoclinally folded sediments of varying composition, the bedding being recognized by alternating bands of quartzite and schist. The strike of the schistosity is consistently north-south to N. 20° E., parallel to the bedding except where there has been intense crumpling.

There are two main schist belts, one east and one west of the Kinsman quartz monzonite. The dips of the schistosity in the belt west of the Kinsman mass vary from vertical to seventy degrees east, except at the eastern contact just west of Mt. Cardigan, where the schistosity has a low dip, averaging twenty degrees to the east. In this marginal zone, pressure from the advancing magma has crumpled the schists and overturned the drag folds toward the west. This local crumpling below, and in front of, the intrusion may be seen at several places along the western side of the summit of Mt. Cardigan and on the adjoining peaks. The width of the belt of overturning and crumpling varies from a few feet to a mile. Drag folds of varying size are found throughout the schist, away from the contact zone. An excellent example of these is seen in exposures opposite Tewksbury Pond, along United States Highway No. 4, where an overturned fold, 15 feet high, is visible in the road-cut. East of the contact in the Fowler River, 2 miles west of the Crawford School, other sizable drag folds can be seen.

East of the Kinsman intrusion the dips of the schistosity are steep, the southern section averaging sixty to eighty degrees west, the northern, seventy degrees to the east.

Along the northern border of the intrusion the strike of the Littleton

schists is variable, although there is a marked tendency for the schistosity to strike northeast, curving around the north end of the Kinsman sill. In this section, where drag folds are well developed, they are overturned to the west, as in the other regions. The plane of the axes of the folds dips about fifty degrees southeast, indicating pressure from that direction. The folds plunge northeast.

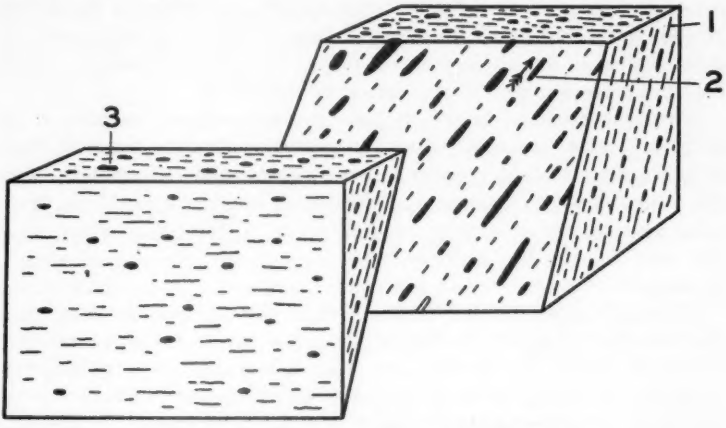

FIGURE 2.—*Lineation as seen on various planes*

Block is split along foliation plane. (1) Foliation; (2) lineation in plane of foliation, arrow indicating direction of flow of magma; (3) mean dimensional elongation in plane of foliation.

There are various isolated pendants of schist in different places throughout the Kinsman quartz monzonite, especially in the south. These parallel the regional strike and dip steeply eastward.

Many of the sillimanite crystals, which are characteristic of the schist, show linear orientation in the plane of schistosity. They do not lie with their long axes in the direction of steepest dip, but at an angle to it. The direction in which the sillimanite is oriented is the linear direction and is parallel to the direction of mass movement (Fig. 2 and Pl. 2).[2] The projected strike of this direction on the surface is approximately N. 60° W., with an average plunge from forty to forty-five degrees southeast.

BETHLEHEM GNEISS

The Bethlehem gneiss has marked foliation, shown by parallelism of biotites and feldspars. The strike of the foliation averages N. 20° E., dipping about forty degrees east. The northern exposures dip eastward

[2] Robert Balk: *Structural and petrologic studies in Dutchess County, New York*, part I, Geol. Soc. Am., Bull., vol. 47 (1936) p. 738-743, 760.

at angles from fifteen to twenty degrees in the vicinity of Thompson Hill, where the Bethlehem gneiss has overridden the Ammonoosuc volcanics. The dip in the southern section steepens perceptibly, averaging from thirty to fifty degrees east. The western contact with the Ammonoosuc volcanics is sharp. Near the contact the gneissoid character is pronounced because of close spacing of the foliation planes. This is also true of the eastern contact with the Littleton schists, especially in the section exposed in Orange Brook, just east of Orange Basin. In some outcrops, it is possible to detect lineation of the micas and the feldspars. As in the schists, the plunge is down, toward the southeast.

Inclusions of schist in the Bethlehem gneiss, aligned parallel to the foliation planes, indicate clearly its intrusive nature. Some of these xenoliths are clean-cut, angular slivers. Others grade imperceptibly into the gneiss without sharp margins, indicating assimilation. These inclusions may be seen at a point one mile north of Tewksbury Pond, along the west side of the highway. Throughout the region northwest of Tewksbury Pond the contact is irregular, a result of lit-par-lit injection.

Although pegmatites are associated with all the igneous rocks, those that owe their origin to the Bethlehem gneiss and lie in the belt of schist east of it deserve special mention. Here, the end stage products of the magma took advantage of joint planes and planes of schistosity in the overlying Littleton schists. The result is an intermittent series of pegmatites, striking N. 20° E., parallel to the schists east of the main contact. In places, there are as many as five or six of these wide sill-like masses parallel to each other.

North of Orange Basin the higher summits of The Pinnacle, Mt. Tug, and Kimball Hill are capped by these dikes for a distance of 6 miles. Where the pegmatites are in flat sheets, they must have followed horizontal joint planes. As a result, the hilltops are capped by quarter-of-a-mile-wide exposures of dikes, which may have been only from 20 to 100 feet thick.

The best place to study the relationship of the pegmatites to the horizontal joint planes, with the resultant capping of the higher summits, is north of Orange Basin on The Pinnacle and Tug Mountain. Pegmatites controlled by vertical planes of schistosity are well exposed at the Ruggles Mine, one mile south of Isinglass Mountain.

From a study of the structural features of the Bethlehem gneiss, it is evident that it is an intrusive igneous body, which forced its way in between the Littleton schists and the Ammonoosuc volcanics, in the northwestern part of the quadrangle. The source of the uprising magmatic material was to the east. The sill-like character of the Bethlehem gneiss is apparent from a study of its foliation and contact relationship with the

Ammonoosuc volcanics, on which it rests at a low angle. Pegmatites are characteristic end-stage products of the magma and inject the overlying Littleton schists in considerable quantity.

KINSMAN QUARTZ MONZONITE

Comparison with the Bethlehem gneiss.—Structurally, the Kinsman quartz monzonite is much like the Bethlehem gneiss. Both are gneissoid, the former having a coarser, more-pronounced porphyritic texture, caused by the larger orthoclase phenocrysts and the granular quartz-feldspar aggregates. The north-south to N. 20°E. foliation, with an eastward dip, is a feature common to both. There is no structural proof that the two are different in age. They are nowhere in contact with each other, although they are separated by a belt of schist only a mile wide in the Tewksbury Pond region. Pegmatites are associated with the final stages of intrusion of both magmas. They are probably granitic intrusions that rose from the same magma chamber and intruded the Littleton schists at approximately the same time.

Linear features.—It is possible to reconstruct in part the method of intrusion of the Kinsman quartz monzonite, by studying the feldspars and the granulated aggregates of various minerals. They tend to lie with their long dimensions parallel, producing a linear element within the foliation plane. The strike of this alignment, projected on a horizontal plane, averages N. 60°W., plunging to the southeast. Viewed on any surface lying at an angle to the plane of the foliation the true lineation is not seen. Care must be taken in studying a horizontal surface not to interpret the mean direction, which strikes N. 20°E., parallel to the schistosity, as the direction of maximum elongation. Figure 2 shows how this lineation appears in different planes. The linear feature is pronounced near the margins of the intrusion and near inclusions of schist. Its projected strike corresponds closely to the projected strike of the sillimanite crystals in the schist. In the gneiss, the streaking was produced partly by orientation of the crystals which were suspended in the moving magma, and partly as a result of granulation as consolidation continued. As the moving mass became viscous, foliation planes were developed, granulation was extreme, and the resulting ribbons of quartz and feldspar were strung out in the shear planes, parallel to the direction of maximum elongation (Pl. 2).

Structural features near the top of Mt. Cardigan.—On the summit of Mt. Cardigan the Kinsman quartz monzonite is massive and granitic, foliation planes being inconspicuous. Close scrutiny reveals flat-lying biotites. There is also a tendency for phenocrysts and granular quartz feldspar aggregates to lie horizontally or to show a dip not exceeding twenty degrees toward the southeast. This rough alignment, or parallel-

ism, is above the contact with the schists on the west side of the summit, and indicates a low dip for the flow direction, the base of the sill being nearly horizontal near the summit of the mountain. The dip of phenocrysts and biotites steepens eastward from the summit, inclinations up to fifty degrees being found near the eastern contact.

Study of the granite dikes and pegmatites in the Kinsman shows a few early dikes which, unlike the later dikes, have been deformed. These early dikes are small and obscure in contrast to the broad bands made by the later aplites and pegmatites, which cut across the Kinsman in many directions. The earlier series show distortion in the form of drag folds whose axes are almost horizontal, for they have been overturned toward the west. They must have been in existence before the intrusive action had ceased and before the incoming of the later dikes. Pressure from the southeast moved the already crystallizing front of the sill, even after the dikes had formed. After movement had ceased, the late dikes were intruded along planes of weakness, formed at angles to the earliest dikes as a result of new stresses.

Mode of intrusion.—The Kinsman quartz monzonite makes up the entire central part of the Cardigan quadrangle. It intruded the folded Littleton schists, parallel to the schistosity, in the form of a great sill, which arched the schists. As the schists at the base of the sill are overturned toward the west, it is clear that intense deformation accompanied the intrusion. Folding of the earlier dikes shows that the front of the sill had begun to crystallize before the intrusive forces had ceased. The plunging of the linear features to the southeast indicates that the source of the igneous material was from that direction.

CONCORD GRANITE

The intrusion of the Concord granite followed that of the Kinsman quartz monzonite. There is little evidence of deformation of the granite as a whole, except in the region between Spectacle Pond and Mt. Crosby, in the northeastern part of the area. In most places the relationship of the Concord granite and the older rocks is of a lit-par-lit nature, making it difficult to draw clear-cut boundary lines.

Foliation is characteristic of the Concord granite on the west side of Newfound Lake, on Sugarloaf Mountain, and south of Bear Mountain. As the foliation is steep and is parallel to the contact plane, it may have been induced as a result of primary flow. If regional deformation had been the cause of its development, it would have been more wide-spread.

A sill-like strip of Concord granite, northwest of Newfound Lake, caps the ridge of which Mt. Crosby is the highest peak. The sill is not as wide as it appears on the map, for the eastern slope of the ridge follows roughly

the dip slope of the sill and the hanging wall has been irregularly eroded. The apparent width of the sill also has been increased by overthrusting. Evidence for this is found in the base of the sill and may be seen on the cliffs east of Spectacle Pond and in the depression between Mt. Crosby and Bald Knob. Drag folds and minor faults below the foot wall are overturned toward the west. The base of the sill is a fine-grained pegmatite, with flow lines parallel to the contact planes. There are many black tourmaline needles and small garnets within the pegmatite.

Between Mt. Crosby and Jewell Hill, an area of muscovite schist is extremely crumpled and injected by numerous dikes of Concord granite. The position of the schist, directly above the Kinsman intrusion, which is not exposed in this region but crops out a few miles north, in the Rumney quadrangle, may explain its extreme distortion. The strike of the schistosity is variable, ranging from N. 20°E. to east-west. The Concord granite locally makes up fifty per cent of the rock, the intrusion, in general, being controlled by the schistosity, although a cross-cutting relationship is common.

Southwest of Mt. Cardigan, for a distance of 10 miles, the Concord granite has irregularly intruded the contact zone of the schist and Kinsman quartz monzonite. Nowhere in this belt does the granite show marked deformation.

In comparison with the older intrusions, the Concord granite is erratic in distribution. Generally, it is non-foliated and cuts across the preexisting structures of the intruded rocks, indicating that the host rocks were not plastic when it was injected. Locally, it is foliated parallel to the contacts of the schists.

There is evidence that during, and after, the intrusion of the Concord granite pressures were great enough to affect the rocks locally. This is true of the sill on Mt. Crosby, where overthrusting and folding have taken place.

METAMORPHISM

The high degree of metamorphism of the rocks of the Cardigan quadrangle may be attributed largely to intrusive activity accompanied by deformation. The association of many large crystals of sillimanite with the entire metamorphic zone where intrusions have been abundant is significant, although no new material need be added to schists to produce sillimanite. The original composition of the schist has been important in determining whether or not sillimanite developed. As the Kinsman quartz monzonite occupies a far larger area than the other two intrusions, and as the Concord granite came in after movement was less pronounced, the Kinsman injection caused most of the metamorphism.

In many places the Littleton schists have been thoroughly soaked by the magmas, so that the boundary between schist and granite is indefinite. This is especially true east of the main contact of the Kinsman quartz monzonite, where granitization of the schists has taken place in belts up to half a mile in width. On Hutchins Hill and Simmons Ledges in Alexandria, where alternate bands of gneiss and schist are characteristic, soaking of the schist has resulted in the development of metacrysts of plagioclase feldspar. Their quantity varies considerably, increasing near roof-pendants and contacts. Their composition is similar to that of the plagioclase in the neighboring bands of igneous rocks.

Evidence for a certain amount of assimilation in the granite is found in the presence of sillimanite needles, although there is no other trace of the schist in which they originated. Sillimanite is also common in the vicinity of roof-pendants, where foliation, resulting from shear and granulation, together with assimilation, were at a maximum development.

Throughout the Littleton schists, garnets of all sizes are abundant, some bands containing more than others, as a result of original differences of composition and proximity to the igneous masses. Unusually large and abundant garnets are associated with the contact of the Kinsman quartz monzonite and the schist. Garnets are found in greatest quantity in the roof-pendants in the southern part of the quadrangle. They average half an inch in diameter at the North Wilmot garnet mine, 4 miles southwest of Danbury, and make up as much as fifty per cent of the rock. Near the contact of the smaller sill, in the southeast section of the quadrangle, garnets averaging one inch in diameter are common. Even larger garnet aggregates, measuring up to three inches, are found near the Kinsman contact in the schist on Taylor Hill, one mile northeast of Danbury, and in the Fowler River just east of the contact, about two miles northwest of the Crawford School. In places, they resemble augen, the granular garnet aggregates making up the central knots, with pegmatitic materials stringing out at either end, and with biotites outlining the whole. These aggregates are typical of the east contact zone, and have not been observed west of the Kinsman intrusion.

SUMMARY OF INTRUSIVE HISTORY

A picture of the intrusive history of the Cardigan quadrangle is obtained by assembling the facts derived from a study of: (1) structural features of the schists and plutonic rocks, (2) age relationships, (3) petrology, and (4) metamorphism.

Both the Bethlehem and the Kinsman gneisses are sill-shaped bodies whose upper surfaces dip more steeply eastward than their lower surfaces. The long narrow belt of Concord granite that makes a subsidiary sill on

Mt. Crosby has the same structure. Proofs of the method of emplacement of these igneous bodies are best understood from a study of the Kinsman porphyritic quartz monzonite, for it has the largest surface distribution of the rocks of the quadrangle.

That the Kinsman gneiss was injected into the Littleton schists from the southeast, in the form of an elongate sill, and that it buoyed the schists up in an arch, overturning them toward the west, is proved by: (1) plunge of the linear features to the southeast; (2) overfolding of the schists to the west, the axial planes of the folds paralleling the foliation of the magmatic materials; (3) details of the structure at the base of the Kinsman mass and the contact zone of the schist; (4) intrusive effects on the schists. These structures are found, though less well developed, in the other intrusions, giving proof of their similar origin.

The relative age of the Bethlehem and the Kinsman gneisses is determined by correlation with rocks of other quadrangles. In the Cardigan quadrangle, their magmatic relationship is well shown, both by their similarity, as seen in the field, and by the study of thin-sections. The younger age of the Concord granite is shown by its cross-cutting character; and its magmatic relationship by the megascopic and the microscopic resemblance of the coarser phase to the older intrusions. The fine-grained phase is gradational into the coarser phase.

The two older sills must have been intruded contemporaneously or in close succession, into previously folded sediments, parallel to the planes of schistosity. Pressure continued until the magmas had almost congealed. Foliation developed by recrystallization and granulation as a primary flow phenomena. By the time the later member of the series, the Concord granite, was intruded, the host rock was less plastic. The magma, therefore, followed complex pre-existing structures, in places cutting across them, resulting in an erratic distribution. The granite shows the effect of the westward pressure best in the sill on Mt. Crosby.

The great amount of pegmatite associated with all three intrusions, and the extent of the lit-par-lit injection, show that in the late stages the magmas were charged with water and volatiles.

At least part of the sillimanite must have developed at a late stage, because its delicate needles penetrate quartz, unbroken by later movement. Unusually large muscovite crystals in the feldspars and interstitial myrmekite also resulted from late hydrothermal activity.

Lack of bending, breaking, or granulation of the minerals formed during late hydrothermal action, and absence of foliation in the latest phase of the Concord granite, show that there has been little movement since final solidification of the magma.

ECONOMIC DEPOSITS

There have been productive mica and feldspar mines in the pegmatites in the vicinity of Orange Basin, Isinglass Mountain, and Prescott Hill, and the region has many abandoned prospects. Sterrett, in 1914, wrote a comprehensive paper on the mica deposits of the United States, describing some of the deposits in the Cardigan region. The most productive mines lie in the belt of pegmatites east of the Bethlehem gneiss. The only large mine ever worked east of this belt is at a point two miles west of Alexandria. The Ruggles mine, one mile south of Isinglass Mountain, is the only sizable mine being worked at present. Here, the pegmatite is made up of quartz and feldspar, biotite and muscovite. Black tourmaline is a common accessory. Among the rarer minerals are beryl, montmorillonite, uraninite, and other radium minerals. Feldspar is the sole mineral being exported at present. The mineralogy of this deposit is being studied by W. E. Richmond, of Harvard University, and B. Shaub, of Smith College. Minor prospects are periodically exploited in various parts of the quadrangle, but none of these produces on a large scale.

The North Wilmot garnet mine, southwest of Danbury, described by Conant, exports garnets. It has such a large reserve on hand, however, that there has been no active mining in recent years.

GEOMORPHOLOGY

The oldest rocks in the Cardigan quadrangle are the Ammonoosuc volcanics, of Ordovician age. They are of various kinds, so highly altered that their igneous origin can scarcely be recognized. They were intruded by the Smarts Mountain granite, of the Oliverian magma series, at the close of Devonian time, before the incoming of the New Hampshire magma series.

The Littleton schists were laid down in Devonian seas, as vast deposits of muds and sands. It is impossible to determine to what extent these sediments had been folded before the incoming of the great igneous intrusions of the New Hampshire series. For, as the Bethlehem and the Kinsman magmas worked their way upward from the southeast, the compression was so great that the sediments were sheared and crumpled. The combined heat and pressure, together with penetrating magmatic materials, metamorphosed them to biotite—sillimanite-garnet schists and quartzites. Granulated masses were elongated in the direction of magmatic flow. The sillimanites in the schists are oriented parallel to this same direction of movement. At the base of the Kinsman, the schists were locally overturned, drag folds developing in the schists throughout, their planes dipping to the southeast. The Concord granite was intruded as the late phase of magmatic activity. Fine-grained dikes were asso-

ciated with the late phase of its intrusion. Pegmatites were given off from the igneous masses as end-stage products. Later still, a small fault developed, offsetting the Ammonoosuc volcanics in the northwest. A few scattered basic dikes of uncertain age have been found.

The more recent history of the region has been one of profound erosion, followed by glaciation. The Cardigan region is maturely dissected, with a rough north-south to northeast-southwest alignment of the topography, parallel to the regional strike. Mt. Cardigan, Tug Mountain, Bear Mountain, Mt. Crosby, and Tenney Mountain show this alignment. The main drainage lines, however, cut across the structure, and appear to have inherited their courses from an earlier erosional surface. The summits are not accordant, so that the evidence for a former erosion level is poor.

Pleistocene glaciers passed over the area, rounding and modifying the outlines of the hills. The direction of movement was from northwest to southeast, as shown by glacial striations and gouges, which strike N. 60° W. Near the summit of Mt. Cardigan, lunoid furrows, a foot or more in diameter, show this same direction of movement. In many places, glacial drift plasters the hillsides, concealing the bedrock. This is particularly true on the western slopes.

Glacial deposits of drift and outwash have blocked the main streams, causing small lakes and swamps, especially in the Smith River region. There was a small lake in the Smith River valley at one time, for as many as five terrace levels in the sands and gravels of the old lake bed are seen east of South Alexandria. Varves are found in the lower part of the gravel sections. The present course of the river, over a series of rapids to the Pemigewasset, is post-glacial, the former course having been through the wide gap east of the Gordon School, to join the Newfound River at Bristol.

During glacial times, Newfound Lake was much larger than it is now. At one time, an outlet south of Bog Brook drained to the Smith River. The lake itself was further deepened, for it lay directly in the path of the ice; the depth at The Ledges is reputed to be 135 feet. Blocking of the outlet by glacial debris and outwash, with later downcutting, took place after the ice recession. Bristol lies in what was formerly the outwash plain.

The many large potholes along the river courses were scoured out by water from melting ice. The most spectacular are those at "Sculptured Rocks," a quarter of a mile west of School No. 9, where a deep gorge is cut in the Kinsman quartz monzonite along the Cockermouth River. Thus, the natural beauties of the Cardigan quadrangle are due to the combined forces of stream dissection and glaciation.

CONCLUSIONS

The study of the rocks of the Cardigan quadrangle shows that sediments of the Littleton formation have been isoclinally folded, injected by a series of sills, and intensely metamorphosed. Most of the evidence for the mechanics of the injection has been derived from the study of the largest of these sills, the Kinsman quartzite monzonite (a porphyritic gneiss), and its contact phenomena. Linear features in this gneiss show that the magma rose from the southeast. As a result of this injection, the schists throughout the region, and particularly those at the base of the Kinsman sill, were overturned toward the west. Lit-par-lit-injected contact zones, in which roof pendants of schist alternate with gneiss, are in some places a mile wide. The high degree of metamorphism of the schists is largely attributed to the magmas, for abundant garnets and a high percentage of sillimanite are contact and regional metamorphic phenomena that characterize the Cardigan quadrangle and are not found west of the area affected by the group of sills.

The various intrusions of the Cardigan area are considered to be members of a single magmatic series, for (1) Field studies show the rock types to be gradational in some places; (2) their similar mode of intrusion suggests close sequence in time of injection; (3) the study of thin sections shows a gradual change in composition and similar textural features within the various rock types, a pure granite representing the last phase of the intrusion; (4) movement within the magmatic bodies had almost ceased by the time the last hydrothermal action of the earlier phases of the intrusion had occurred, the injection of the youngest rocks of the series being simultaneous with this late action.

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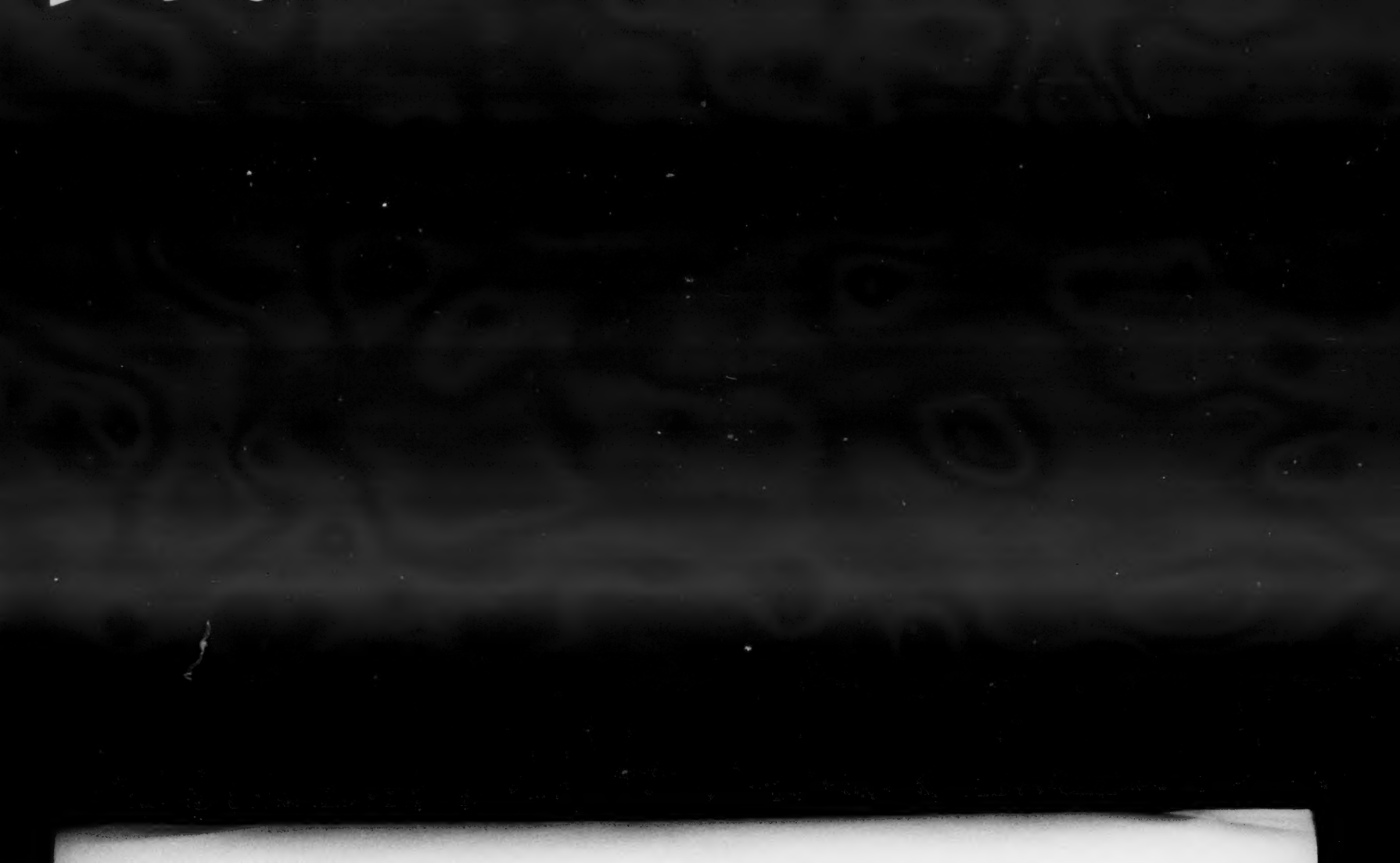

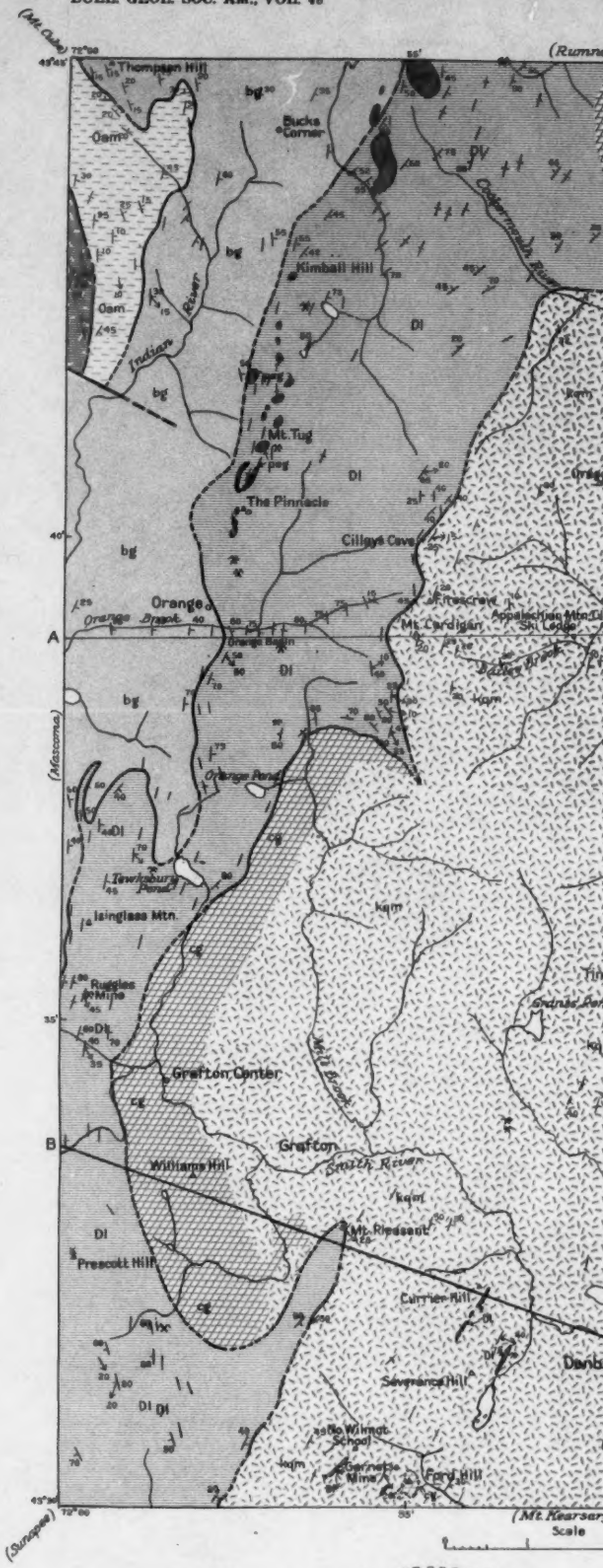

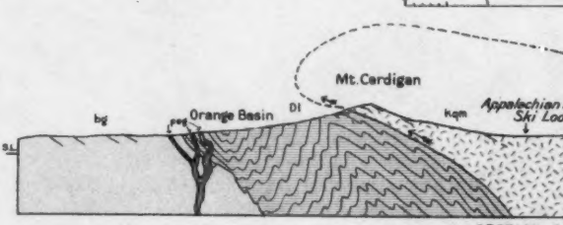

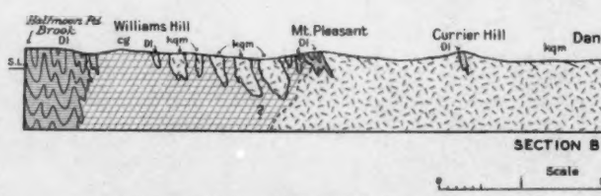

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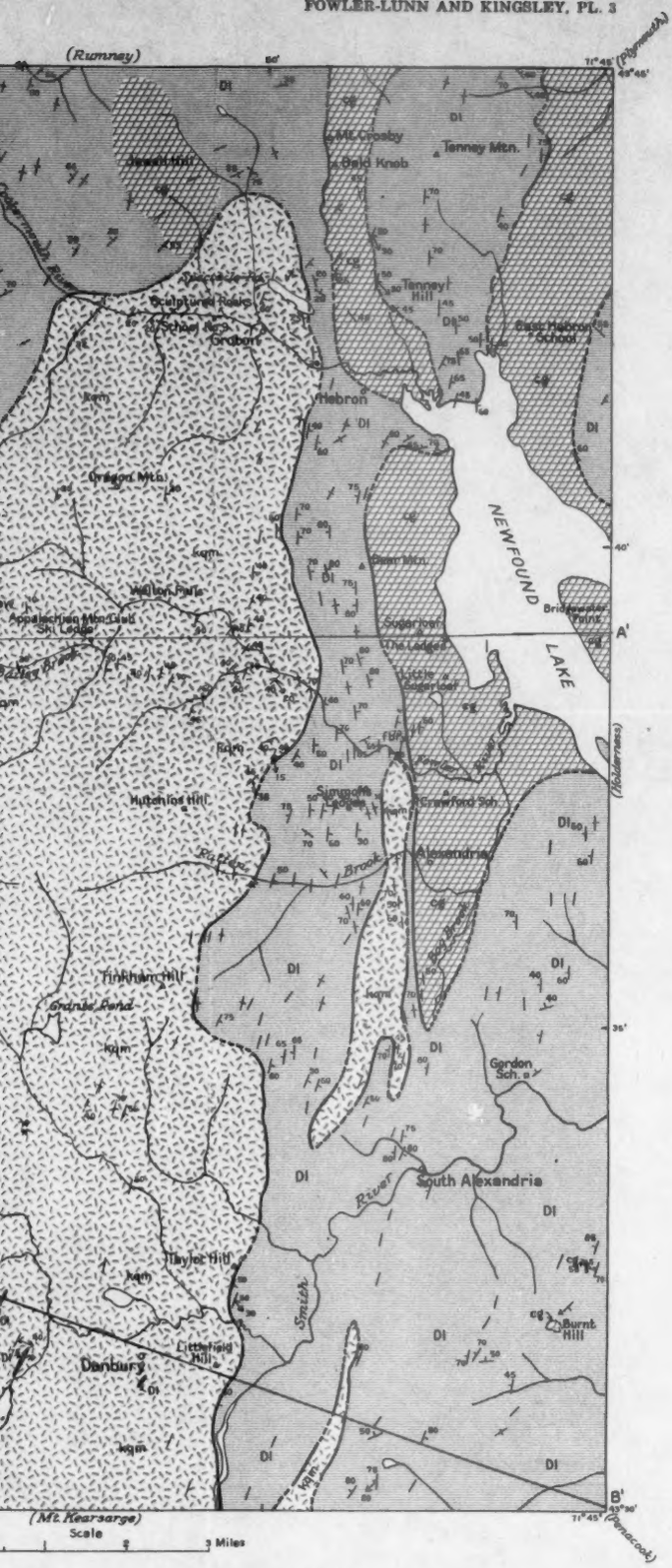
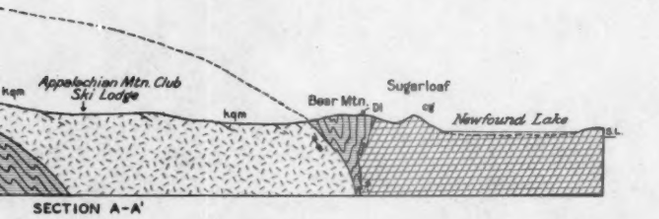


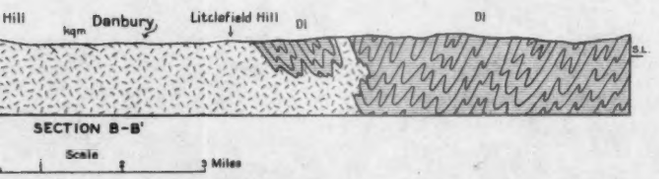
LEGEND

NEW HAMPSHIRE MAGMA SERIES
Late Devonian and possibly Carboniferous

- fbr — "Franconia" breccia (*Osterops only, not a mapable unit*)
- Cg — Concord granite
- kqmp — Kinsman quartz monzonite
- Pegmatites
- bg — Bethlehem gneiss
- Ol — DEVONIAN--Oliverian magma series (*Smarts Mountain granite*)
- Dl — DEVONIAN--Littleton schists
- Oam — ORDOVICIAN--Ammonoosuc volcanics
- Mines or sizeable prospects
- Boundaries of formations
- Inferred formational boundaries
- Fault
- Strike and dip of schistosity
- Strike of schistosity where dip is undetermined
- Vertical schistosity
- Lineation with pitch
- Strike and dip of schistosity with lineation and pitch
- Direction of flow of magma



SECTION A-A'



SECTION B-B'

Scale
3 Miles

OF THE CARDIGAN QUADRANGLE, WITH CROSS SECTIONS

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EDSEL FORD RANGE
Air-photo of central portion of the range.

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PETROLOGIC AND STRUCTURAL RELATIONS OF THE EDSEL FORD RANGE, MARIE BYRD LAND, TO OTHER ANTARCTIC MOUNTAINS

BY F. ALTON WADE

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INTRODUCTION

The Edsel Ford Range is located in northwest Marie Byrd Land. It extends for approximately 150 miles along the eastern boundary of Sulzberger Bay and acts as a retaining wall to check the advance of ice from the Rockefeller Plateau. The geographic setting is shown in Figure 1. This section of the Antarctic was discovered in 1929 by Admiral Byrd during an exploratory flight that carried his plane to within about 100 miles of the range. A reconnaissance map of the area was constructed from the photographs taken during the flight.[1] A four-man sledge party led by Paul Siple, which included in its personnel Stephenson Corey, Olin Stancliff, and the writer, spent 39 days of the short field season in these mountains. From the observations made and from the results of the petrographic analyses of the rocks collected, comparisons have been made and relationships of the Edsel Ford Range with the known regions of the Antarctic have been tentatively established.

[1] E. Byrd: *The flight to Marie Byrd Land*, Geogr. Rev., vol. 23 (1933).

STRUCTURE OF THE EDSEL FORD RANGE

PHYSIOGRAPHY

The Edsel Ford Range is composed of a series of mountain groups whose present configurations are the combined results of the folding of a series of sedimentary formations, the intrusions of great masses of igneous rocks, and the erosive action of local glaciers and the continental ice sheet (Pl. 1). Considered individually, the mountain groups composing this range present varied and dissimilar physiographic expressions. Those composed largely of folded metamorphosed sediments are usually somewhat pyramidal and the peaks of such groups are connected by snow-covered ridges. Mt. Saunders and the Raymond Fosdick Mountains have serrated outlines. Mt. Rea is a cluster of peaks with steep faces and flat tops. In most cases two factors have predominated in the production of the present shapes: first, the attitude of the major joint systems, and, second, the erosive action of local glaciers. The tremendous force produced by water freezing in joints has caused the splitting off of great blocks of rock which are now piled up at the base of the cliffs or have sunken out of sight in the ice. Where the dip of the major joints is nearly vertical, as it is at Mt. Grace McKinley and at Mt. Rea, this action has produced smooth, steep cliffs. More gentle dips result in more gentle slopes.

The great valleys separating the various mountain groups owe their existence principally to zones or planes of weakness in which erosive action has been greatest. Major joint systems are essentially parallel to each of the indentations in the hypothetical coast line. As these valleys are the principal outlets to the sea, they have been greatly deepened by the ice moving down them.

The sedimentary formations are intensely folded and the beds at most of the observed outcrops are nearer vertical than horizontal. However, in the Chester Mountains, the beds of slate composing one small peak are nearly flat-lying and relatively undisturbed (Pl. 2, fig. 1), while the beds of the surrounding peaks are greatly disturbed.

The major fold axes are not parallel. There appears to be a spreading of the axes; those to the south of John Hays Hammond Inlet extend in a direction parallel with King Edward VII Land Peninsula[2] and those to the north of the Inlet swing northward (Fig. 1). This folded area was intruded by great masses of granite and granodiorite. A contact of the metamorphosed sediments and the intrusives is exposed on the northernmost peak of Mt. Rea. Great fingers of granite intrude the black schists, cross-cutting structure in the main (Pl. 2, fig. 2). Stoped blocks

[2] F. A. Wade: *Northeastern borderlands of the Ross Sea: glaciological studies in King Edward VII Land and northwestern Marie Byrd Land*, Geogr. Rev., vol. 27 (1937).

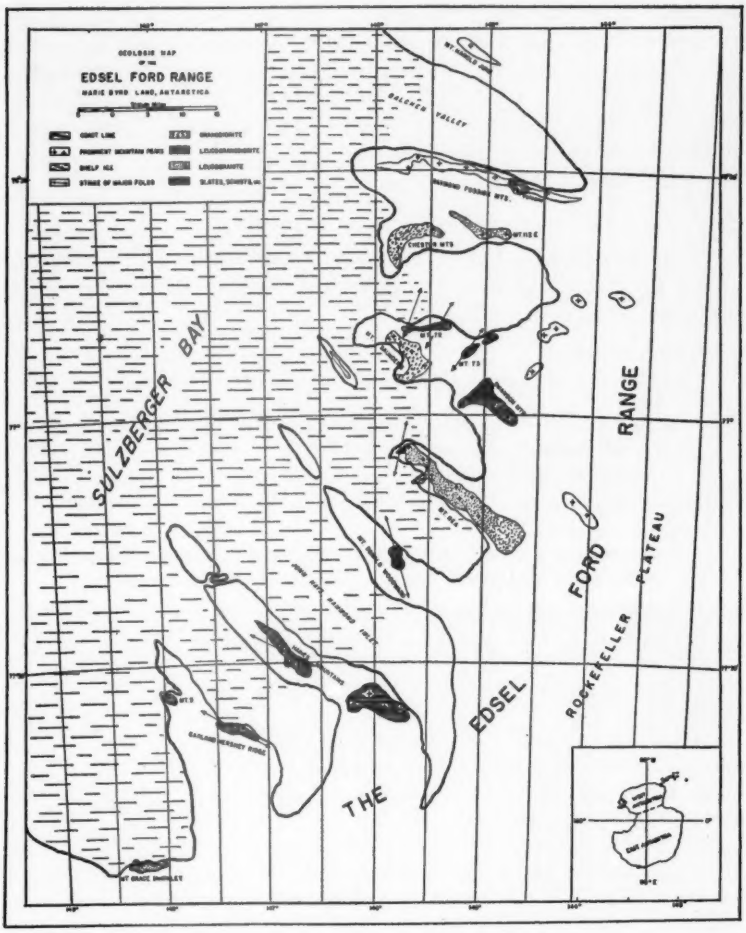

FIGURE 1.—*Geologic map of the Edsel Ford Range*

of the schist can also be seen high up on the face of a nearby granite peak (Pl. 3, fig. 1).

The physiographic expression of the range as a whole resembles the outlying foothills of the Queen Maud Range,[3] but certainly the tabular-topped mountains, characteristic of this and other ranges of the great

[3] L. M. Gould: *Structure of the Queen Maud Mountains, Antarctica*, Geol. Soc. Am., Bull. 46 (1935) p. 974.

Antarctic Horst, are not comparable. The Queen Maud foothills are composed of pre-Cambrian gneisses, schists, and intrusive granites and owe their present forms to frost action and the erosive action of moving ice. The similarity stops here, for the series of younger folded sediments is entirely lacking and there is no evidence of the pre-existence in the Edsel Ford Range of a series of sediments similar to the Beacon sandstone series which formerly capped the Queen Maud foothills.

PETROLOGIC SUMMARY

The rocks composing the mountains of the Edsel Ford Range as represented by the specimens collected by the Eastern Sledge party may conveniently be grouped as follows:

I. Igneous rocks
   a. Massive intrusives. Granodiorite, leuco-sodaclase granodiorite, leucogranite
   b. Dike rocks. Porphyritic diabase, diabasic diorite, granite porphyry, granodiorite, leuco-sodaclase granodiorite, amphibole peridotite
   c. Extrusive. Olivine basalt

II. Metamorphic rocks
   a. Granite gneiss
   b. Metamorphosed sediments. Biotite schist, orthoclase-biotite schist, biotite-orthoclase schist, sericite schist, orthoclase-sericite schist, slate

III. Sedimentary rock. Arkose

The following were collected as erratics: leuco-sodaclase granodiorite, granodiorite, leucogranite, olivine basalt, slate, phyllite, and biotite-muscovite schist.

In Table 1 are listed the mineralogical analyses of the massive intrusives and in Table 2 those of the principal dike rocks.

With but few exceptions the igneous rocks of this area are high in sodium and potassium. The small percentages of mafic minerals are especially worthy of note. In only one massive intrusif (Mt. Saunders) is a pyrabole an essential constituent. The plagioclases of the granites and granodiorites are commonly zoned (Pl. 3, fig. 2). The olivine basalt is from an extinct volcano located on the southern slope of the Raymond Fosdick Mountains (Fig. 1). From the relations of the extruded ash and fragmental material to the ice, it is evident that the eruption continued after the inundation of the region by ice.

Of the metamorphic rocks observed and studied the only one which might be considered to represent the basement complex of East Antarctica[4] is the granite gneiss of the Raymond Fosdick Mountains. The

[4] *Op. cit.*, p. 980.

FIGURE 1. FLAT-LYING SEDIMENTARY FORMATIONS
In Chester Mountains. Dip of beds in the peak in right background is approximately 50 degrees.

FIGURE 2. LEUCOGRANITE STRINGERS
In schists at Mt. Rea.

FIGURE 1. STOPED BLOCK OF BLACK SCHIST
In granite of Mt. Rea.

FIGURE 2. ZONED PLAGIOCLASE
In Mt. Rea granite. Crossed nicols, X 53.

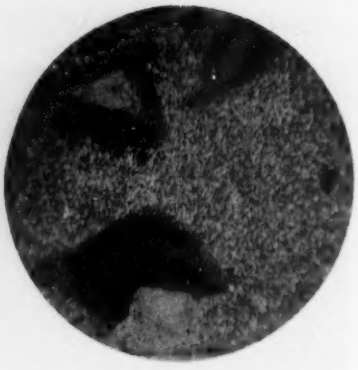

FIGURE 3. ORTHOCLASE-SERICITE-SIDERITE SCHIST
Haines Mountains. Siderite partly altered to limonite. Crossed nicols, X 20.

metamorphosed sedimentary series has not undergone such intense alterations as the basement gneiss, and no minerals characteristic of high temperature metamorphism were noted in the thin sections examined. Recrystallization varies in intensity, and cataclastic texture is excep-

TABLE 1.—*Mineralogical composition of the Massive Intrusives of the Edsel Ford Range of Marie Byrd Land, Antarctica*

Mineral	Specimen									
	1	2	3	4	5	6	7	8	9	10
Quartz	21.8	41.5	31.5	21.3	26.2	13.6	7.2	18.8	5.1	10.6
Orthoclase	18.7	44.2	49.9	.....	13.3	18.0	15.1	24.6	15.4	24.8
Microperthite	.....	.....	.....	65.5	.....	.....	.....	3.9	.....	43.3
Albite	56.6	.....	.....	.....	.....	.....	.....	.....	.....	.....
Albiclase	.....	.....	.....	.....	.....	.....	68.6	.....	.....	18.5
Oligoclase	.....	.....	.....	9.8	.....	.....	.....	.....	.....	.....
Andesine	.....	7.7	16.8	.....	40.7	48.8	.....	41.7	58.9	.....
Muscovite	2.8	.....	p	1.7	.....	.....	0.5	.....	.....	.....
Biotite	.....	6.7	1.7	1.7	17.3	17.9	8.5	10.9	14.3	2.7
Hornblende	.....	.....	.....	.....	2.4	1.7	.....	.....	.....	.....
Magnetite	.....	p	p	.....	p	p	.....	.....	4.9	p
Apatite	p	p	p	p	p	p	p	p	1.4	p
Zircon	.....	.....	p	p	p	p	p	p	p	p
Beryl	p	.....	.....	.....	.....	.....	.....	.....	.....	.....
Topaz	p	.....	.....	.....	.....	.....	.....	.....	.....	.....
Calcite	.....	.....	.....	.....	p	p	.....	.....	.....	.....
	99.9	100.1	99.9	100.0	99.9	100.0	99.9	99.9	100.0	9.99

p=present in thin section.
1. Leuco-sodaclase granodiorite. Mt. Grace McKinley.
2. Leucogranite. Mt. Grace McKinley.
3. Leucogranite. Mt. Grace McKinley.
4. Leucogranite. Mt. Rea.
5. Granodiorite. Mt. Saunders.
6. Granodiorite. Mt. Saunders.
7. Sodaclase granodiorite. Fosdick Mountains.
8. Sodaclase granodiorite. Chester Mountains.
9. Granodiorite. Fosdick Mountains.
10. Leucogranite. Chester Mountains.

tional. One bed of arkose has been only very slightly altered. The outstanding mineralogic characteristic of the schistose rocks is the prevalence of feldspar over quartz, indicating that the original sediments had suffered little decomposition. The presence of siderite (Pl. 3, fig. 3) is noteworthy. No fossils were found.

From the rather meager data it is concluded that the sequence of events was as follows: deposition of a great series of arkosic sandstones

and shales on the pre-Cambrian basement rocks, close folding of this sedimentary series accompanied by the deep-seated intrusion of acid

TABLE 2.—*Mineralogical composition of the dike rocks from the Edsel Ford Range of Marie Byrd Land, Antarctica*

Mineral	Specimen						
	1	2	3	4	5	6	7
Quartz	.....	10.0	.....	13.2	.....	15.9	9.3
Orthoclase	.....	30.0	2.4	27.7	.....	20.0	24.7
Oligoclase	.....	.....	.....	51.5	.....	.....	.....
Andesine	.....	.....	50.6	.....	.....	55.5	42.0
Labradorite	57.4	.....	.....	.....	4.0	.....	.....
Bytownite	.....	40.0	.....	.....	.....	.....	.....
Biotite	.....	20.0	1.8	7.6	.....	10.6	24.0
Chlorite	23.0	.....	.....	.....	.....	.....	.....
Muscovite	.....	.....	.....	p	.....	.....	.....
Augite	8.6	.....	.....	.....	.....	.....	.....
Hornblende	.....	.....	45.2	.....	88.0	.....	.....
Olivine	.....	.....	.....	.....	8.0	.....	.....
Magnetite	9.8	.....	p	.....	p	p	p
Apatite	p	.....	p	p	p	p	p
Zircon	.....	.....	.....	p	.....	p	p
Calcite	1.0	.....	.....	.....	.....	.....	.....
Hematite	.....	.....	p	.....	p	.....	.....
	99.8	100.0	100.0	100.0	100.0	100.0	100.0

p=present in thin section.
1. Porphyritic diabase. Mt. Grace McKinley.
2. Melagrangabbro. Mt. Grace McKinley.
3. Dibasic diorite. Mt. Rea.
4. Granodiorite. Fosdick Mountains.
5. Amphibole peridotite. Fosdick Mountains.
6. Granodiorite. Mt. Saunders.
7. Granodiorite. Mt. Saunders.

magma, a long period of erosion, glaciation, and the extrusion of olivine basalt during the Pleistocene.

RELATION OF THE EDSEL FORD RANGE TO THE REST OF THE ANTARCTIC CONTINENT

Antarctica has long been divided into two natural subdivisions. Structure and petrography are at great variance with each other in the two areas. East Antarctica embraces all of that vast region lying to the south of the Atlantic and Indian oceans and extending to an imaginary line connecting the Ross and Weddell seas. West Antarctica has been largely limited to the island province south of South America now known as the Antarctic Archipelago (*See* Figs. 1 and 2). The extensive and until recently unexplored area between the two has been variously

classified by students of Antarctic geology.[5] Northwest Marie Byrd Land with its Edsel Ford Range is located in this area, and the investigation of these mountains was undertaken primarily to secure information to aid in fixing the position of the line dividing these structural and petrographic units.

Physiographically, West Antarctica is an almost mirror image of Tierra del Fuego. Structurally and petrographically they are also very similar. The intrusive gabbros, granodiorites, and granites of this area are similar in structure and composition to those of the Andes.[6] The Mesozoic sediments were folded from west to east in late Tertiary time.[7] Briefly, then, the principal characteristics of West Antarctica are (1) eruptive rocks of the "Pacific type" and (2) extensive folds.

Structurally and petrographically East Antarctica is very different. The basement complex is composed of pre-Cambrian gneisses, schists, limestones, and graywackes, overlain by a tremendous series of nearly flat-lying Paleozoic and early Mesozoic sandstones, arkoses, and shales.[8] Near the close of the Mesozoic East Antarctica was subjected to a period of block faulting.[9] The great Antarctic Horst which forms the western boundary of the Ross Sea was formed at that time. A period of volcanic activity accompanied and followed these tectonic disturbances and is continuing even today (Mt. Erebus). There is no evidence of Antarctandean folds in this area. The petrographic investigations of Prior,[10] Stewart,[11] and others have shown that the igneous rocks are of the so-called "Atlantic type." Briefly, then, we may think of East Antarctica as a faulted plateau region whose rocks are characterized by a relatively high content of sodium and potassium.

The only areas in this great expanse of land between the Ross Sea and the Antarctic Archipelago which had been examined by field parties prior to the second Byrd Antarctic Expedition were Scott's Nunataks and the Rockefeller Mountains, both in King Edward VII Land (Fig. 2). From his petrographic investigations of the rocks from Scott's Nunataks, Schetelig[12] concludes, "The resemblance of the rock-complex from Scott's Nunataks to the pre-Gondwana basement rock-complex of South Victoria

[5] Griffith Taylor: *Antarctic adventure and research* (1930) p. 89, New York.
L. M. Gould: *op. cit.*, p. 982-983.
[6] J. G. Anderson: *On the principal geological results of the Swedish Antarctic Expedition*, 10th Congr. geol. intern., Mexico, C. R. (1906) p. 729.
[7] Griffith Taylor: *op. cit.*, p. 101.
[8] T. W. E. David: *Antarctica and some of its problems*, Geogr. Jour., vol. 43 (1914) p. 624-626.
[9] L. M. Gould: *op. cit.*, p. 979.
[10] G. T. Prior: *Report on rock specimens, etc.*, Natl. Antarctica Expedition, 1901-1904, Natural History, vol. 1, Geology.
[11] Duncan Stewart, Jr.: *A contribution to Antarctic petrography*, Jour. Geol., vol. 42 (1934) p. 550.
[12] J. Schetelig: *Report on the rock specimens collected on Roald Amundsen's South Pole expedition*, Videnskr.-Selsk. Christiania, Skrifter I, Math.-Nat. Klasse (1915) p. 1-31.

Land with gneisses and intrusive granites seem to show the close connection of King Edward Land to South Victoria Land." Stewart[18] states, ". . . after comparative studies of Gould's specimens from King Edward VII Land with the specimens from the Queen Maud Range of

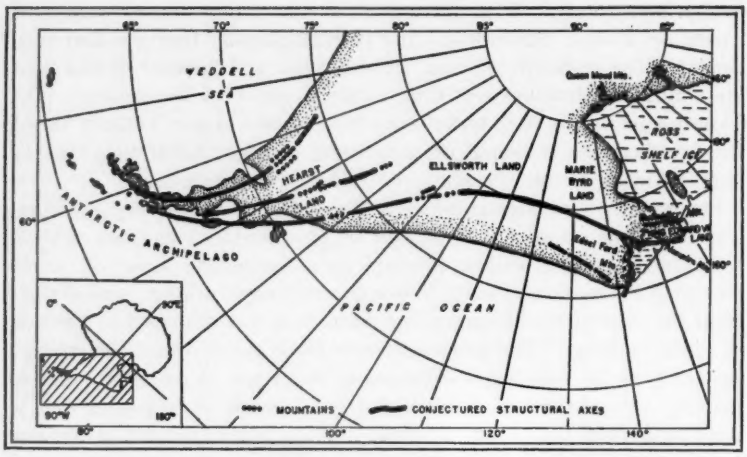

FIGURE 2.—*Conjectured structural axes of West Antarctica*
(In part, after Joerg)

South Victoria Land and with rocks collected in the Antarctic Archipelago by the Expedition Antarctique Française, 1903-05, it is concluded that the rocks of the Rockefeller Mountains have close affinities with the high sodium- and potassium-bearing rocks of East Antarctica, and show little affinity with the high calcium-, magnesium-, and iron-bearing rocks of the Andes of South America and the Antarctic Archipelago."

The northwestern part of Marie Byrd Land has now been added to the known areas. This reduces by several hundreds of miles the length of the unknown in this very important region about which there has been so much speculation. Our new knowledge does not entirely answer the questions, but does give valuable hints as to the solutions of the problems.

The eruptives of northwest Marie Byrd Land are high in sodium and potassium and are in this respect analogous to those of East Antarctica. However, the presence of zoned plagioclases in many of the intrusives suggests a relationship with the rocks of West Antarctica. Structurally

[18] Duncan Stewart, Jr.: *op. cit.*

the area is more like West Antarctica, for it is a region of intense folding and massive intrusions.

In November 1935, Ellsworth definitely established the continuity of land from Hearst Land to the Ross Sea.[14] South of Stefansson Strait he discovered three mountain ranges. Joerg[15] has analyzed the results of the Ellsworth flights and has suggested that these "results seem to confirm not only the conjectured southern continuation of the Antarctandes but also the virgation, or spreading, of their axes forecast by the Swiss geologist Staub"[16] (Fig. 2).

In the light of this new knowledge it seems probable that the Edsel Ford Mountains are a continuation of the northernmost of the three axes. This axis seems in turn to divide in the Edsel Ford Mountains, one of the sub-axes trending northward and the other to the northwest. The latter possibly continues on the eastern side of Sulzberger Bay as the mountains of King Edward VII Land.

The petrographic evidence does not appear to support fully these deductions as the rocks apparently have closer affinities with those of East Antarctica. The granitic intrusives of East Antarctica have been placed in the pre-Cambrian.[17] The age of the granite-granodiorite intrusions into Marie Byrd Land has not been definitely determined, but they have followed the period of folding of the sedimentary formations which certainly must be younger than pre-Cambrian. Correlation between the two is therefore impractical. If the Edsel Ford Mountains are considered to be one of the blind ends of the virgated Andean system, the intrusives might then be considered as high sodic and potassic differentiates of the same magma which was intruded farther to the east in the more central portion of the range.

The true solution of this problem will be known only after considerably more field work has been done in the newly discovered mountains of Hearst and Ellsworth lands and in the Edsel Ford Range. The problem is a most interesting one to both petrographers and structural geologists. Further exploration in these areas should indeed prove to be well worth while.

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[14] Lincoln Ellsworth: *My flight across Antarctica*, Nat. Geogr. Mag., vol. 70 (1936) p. 7-26.

[15] W. L. G. Joerg: *The topographical results of Ellsworth's trans-Antarctic flight of 1935*, Geogr. Rev., vol. 26 (1936) p. 454-462.

[16] Rudolf Staub: *Der Bewegungsmechanismus der Erde*, (1928) p. 117-118; map in pocket, Berlin.

[17] Griffith Taylor: *op. cit.*, p. 106.

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PETROLOGY OF THE SAN MARCOS GABBRO, SOUTHERN CALIFORNIA

BY FRANKLIN S. MILLER

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INTRODUCTION AND ACKNOWLEDGMENTS

This paper includes the results of a petrologic investigation of the gabbroid rocks of the San Luis Rey Quadrangle, Southern California (Fig. 1). The unusual features of these rocks were recognized by Professor E. S. Larsen during his mapping of the quadrangle, and the

present study was undertaken at his suggestion for the purpose of throwing light on their origin and magmatic history.

It is a pleasure for the writer to express his grateful appreciation to Professor Charles Palache and the Holden Fund of the Department of Mineralogy, Harvard University, for financial assistance in connection with both field and laboratory expenses, and to Professor Larsen for his helpful guidance and criticism throughout the investigation. The laboratory work was carried out during the year 1933-1934 while the author was holder of the Woodworth Memorial Fellowship.

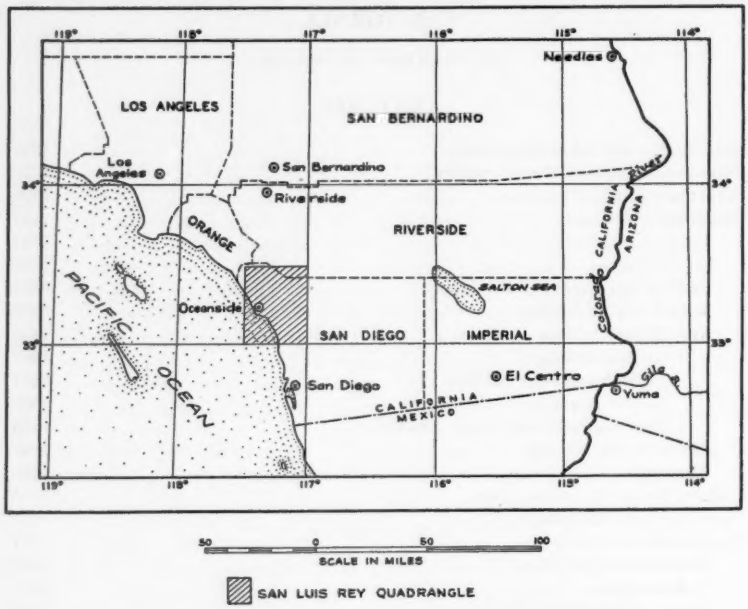


FIGURE 1.—*Location of the San Luis Rey quadrangle*

A composite batholith of Jurassic age forms a large part of the Peninsular Range of Southern California. This batholith intrudes folded Triassic schists and quartzites of sedimentary origin and Jurassic (?) volcanic rocks. In the San Luis Rey quadrangle the batholithic rocks comprise gabbros, tonalites, granodiorites and subordinate granites. The intrusive sequence has progressed normally from the earlier mafic to the later felsic rocks. The gabbros representing the initial phase of this series of intrusions are grouped together as the San Marcos gabbro. Of the tonalites the Bonsall tonalite has been described previously by Hurl-

but,[1] while the Green Valley tonalite will be mentioned below because of its close affinity to the more felsic varieties of gabbro. The most widespread of the granodiorites is called by Professor Larsen the Woodson Mountain granodiorite.

DEFINITION OF THE SAN MARCOS GABBRO

The rocks included in the San Marcos gabbro cover a wide range of composition and texture within the gabbro family. Yet there is almost a complete gradation between the principal rock types, and these types are so intimately associated in the field that it is not practicable to map them as separate units. Moreover, they possess a number of common features which emphasize the close genetic relationship of all the gabbroid rocks. Among these are: similarity in character of outcrops and local structural features; similar relations to the other rocks of the batholith, particularly with regard to age; susceptibility of all the hornblende-free rocks to similar local or widespread enrichment in hornblende; noteworthy similarities in mineralogy; and systematic changes in mineral content and chemical composition throughout the rocks of the group.

The most abundant rock type, and one which represents about the average of the group in composition and texture, is a medium-grained dark-gray norite. Locally the norite grades to hypersthene gabbro. This norite type makes up the greater part of the San Marcos Mountains, from which the name San Marcos gabbro is derived. The other two principal hornblende-poor types are, on the one hand, the olivine norites and olivine gabbros (eucrites), and on the other, those norites and hypersthene gabbros carrying small amounts of quartz and biotite. The threefold division is emphasized by systematic differences in the composition of the plagioclase. In addition to the hornblende-poor rocks, the formation includes a large group of hornblende gabbros nearly equal in areal extent to all the hornblende-poor rocks together. From the standpoint of plagioclase composition the hornblende gabbros fall into three groups which correspond respectively to the norites, olivine norites, and quartz-biotite norites. The primary division of the gabbroid rocks is thus based on the distribution of olivine versus quartz rather than on the abundance or paucity of hornblende, and the key to the interpretation of the rock differences within the formation is found in the variable composition of the plagioclase feldspar.

It is characteristic of the entire group of rocks that wherever pyroxene is present, hypersthene is prominent, and usually predominant. How-

[1] C. S. Hurlbut: *Dark inclusions in a tonalite of Southern California*, Am. Mineral., vol. 20 (1935) p. 609-630.

ever, the presence of considerable amounts of hornblende gabbro, hypersthene gabbro, and olivine gabbro renders the term norite unsuitable for the formation as a whole.

OCCURRENCE AND FIELD RELATIONS

The San Marcos rocks occur in a number of small bodies scattered among the larger tonalite and granodiorite masses of the batholith (Fig. 2). Within the San Luis Rey quadrangle some 40 gabbro bodies occupy a total area of about 60 square miles. The largest single body, which includes the San Marcos Mountains, covers 11 square miles, and nearly a score of other major bodies have areas of 1 to 5 square miles. How far beyond the limits of this quadrangle the batholithic rocks extend without essential change of character is not definitely known. Gabbros apparently identical with the San Marcos types are present among the plutonic rocks of the Cuyamaca region [2] and the adjoining La Jolla [3] and San Jacinto quadrangles.[4] The orbicular gabbro at Dehesa [5] also corresponds to an exceptional phase of the San Marcos gabbro.

The field relations of the gabbros show locally that they are intrusive into the metamorphosed Triassic sediments. The more silicic rocks of the batholith, that is the tonalites and granodiorites, are in turn intrusive into the gabbros and frequently contain abundant gabbroid inclusions, as well as structural features indicative of their intrusive relationship.[6] Thus the San Marcos rocks are the earliest rocks of the batholith and presumably must have occupied a far greater proportion of the batholithic area, before they were intruded and in part assimilated by the succeeding more silicic magmas.

Of the batholithic rocks the gabbros are notably more resistant to weathering than the tonalites and slightly more resistant than the granodiorites, with the possible exception of the Woodson Mountain granodiorite. Hence most of the gabbro bodies stand up as prominent hills in the present topography. On their outcrops the gabbros commonly show marked variations in appearance within short distances. These apparent variations in most cases represent considerable actual differences in mineral content, grain size, and texture within the same gabbro

[2] F. S. Hudson: *Geology of the Cuyamaca region of California*, Univ. Calif., Dept. Geol. Sci., Bull., vol. 13 (1922) p. 175-252.

[3] M. A. Hanna: *Geology of the La Jolla Quadrangle, California*, Univ. Calif., Dept. Geol. Sci., Bull., vol. 16 (1926) p. 187-246.

[4] D. M. Fraser: *Geology of the San Jacinto quadrangle south of San Gorgonio Pass, California*, Mining in Calif., vol. 27 (1931) p. 494-540.

[5] A. C. Lawson: *The orbicular gabbro at Dehesa, San Diego County, California*, Univ. Calif., Dept. Geol. Sci., Bull., vol. 3 (1904) p. 383-396.

H. H. Kessler and W. R. Hamilton: *The orbicular gabbro of Dehesa, California*, Am. Geol., vol. 34 (1904) p. 133-140.

[6] C. S. Hurlbut: *op. cit.*, p. 618.

body. The most widespread and marked type of variation results from sudden changes in the amount or character of the hornblende. At other places the rock appears to be made up of two distinct phases, with one

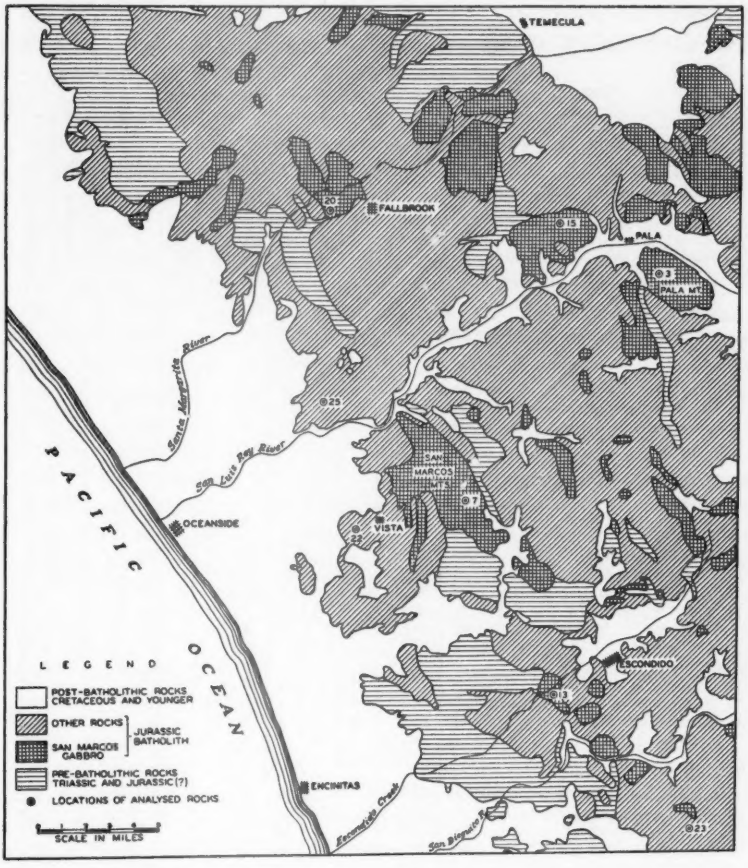


FIGURE 2.—*Distribution of the San Marcos rocks within the San Luis Rey quadrangle*

intimately penetrating the other. Banded structures occur in several localities, and there are exposures of a well-developed nodular gabbro and an orbicular gabbro. These local structures will be discussed in a later paper.

PETROGRAPHIC FEATURES

GENERAL STATEMENT

The principal variations in mineral content within the San Marcos rocks are shown in Table 1. The estimated modes of representative specimens are given, as well as the average composition of each of the main gabbro types. These estimates are based on a number of precise

TABLE 1.—*Modes of representative San Marcos rocks*

	% An in Plagioclase	Plagioclase	Potash feldspar	Quartz	Biotite	Hornblende	Augite	Hypersthene	Olivine	Iron ore	Spinel	Iddingsite	Chlorite
OLIVINE ROCKS													
1 Troctolite (allivalite)	95	75	...	...	...	3	....	1	20	.1	.5	.1	....
2 Olivine norite	93	66	...	...	...	13	....	11	9	.3	.5	.1	....
*3 Olivine norite (with hbd)	93	71	...	...	...	10	....	2	15	.1	1.5	.2	....
4 Olivine gabbro (eucrite?)	94	70	...	...	...	10	5	....	12	2.	.7	.2	....
5 Hornblende-olivine gabbro	89	60	...	...	...	21	1	....	16	.5	1.	.1	....
6 Anorthosite	90	95	...	...	...	4	1	....	....	.2	....	....	....
Average	93	70	...	...	...	10	1	4	14	.3	1.	.1	....
NORITES AND HYPERSTHENE GABBROS													
*7 Norite	64	70	...	...	...	1	3	25	....	1.	....	....	....
8 Norite	60	68	...	...	...	6	8	18	....	.5	....	....	....
9 Hypersthene gabbro	62	71	...	...	...	4	17	6	....	2.	....	....	....
10 Hypersthene gabbro	55	60	...	...	...	15	13	11	....	1.	....	....	....
11 Porphyritic hypersthene gabbro	59	66	...	...	...	3	17	11	....	3.	....	....	....
12 Norite (with hbd)	57	68	...	...	...	12	4	14	....	2.	....	....	....
Average	61	68	...	...	...	5	7	19	....	1.	....	....	....
QUARTZ-BIOTITE NORITES													
*13 Norite (with hbd, qtz, bi)	59	67	...	1	1	12	...	14	...	....	5.	....	....	....
14 Quarts-biotite-hypersthene gabbro	52	68	...	3	1	4	13	10	....	1.3	....	....	....
*15 Quarts-biotite norite	48	64	...	8	6	1	5	14	....	2.	....	....	....
Average	53	66	...	3	3	4	8	14	....	1.5	....	....	....
HORNBLENDE GABBROS													
16 Calcic hornblende gabbro	93	64	...	...	...	23	8	3	....	1.	....	.1	1.
17 Calcic hornblende gabbro	90	65	...	...	...	31	....	....	....	.1	....	....	4.
18 Hornblende-anorthite dike	90	42	...	...	...	50	....	1	...	....	7.	....	....	....
19 Noritic hornblende gabbro	64	61	...	...	.1	27	3	7	....	1.5	....	....	....
*20 Noritic hornblende gabbro	57	51	...	...	...	42	....	5	...	....	1.1	....	....	.2
21 Quarts-biotite-hornblende gabbro	46	56	...	5	8	18	6	5	....	2.	....	....	....
*22 Quarts-biotite-hornblende gabbro	52	58	...	17	7	9	....	7	...	....	1.5	....	....	....
Average	....	58	...	1	1	30	3	4	....	2.	....	....	.5
TONALITES (for Comparison)													
*23 Green Valley tonalite	42	51	...	26	12	8	....	3	...	....	.5	....	....	....
24 Average Green Valley tonalite	40	58	3	18	8	9	....	3	...	....	1.	....	....	....
*25 Bonsall tonalite	41	51	4	26	13	6	....	....	....	....	....	....	....	....

* Analyzed rock, *see* Table 2.

Rosiwal determinations. They are stated in volume percentages, and the same is true of the other statements of mineral quantity throughout the paper, except those in Table 2, which have been recalculated to weight percentages for purposes of direct comparison. In the following brief descriptions, emphasis is laid on those textural and mineralogic features which are significant with regard to the origin of the rocks.

OLIVINE ROCKS

Although not the most abundant of the gabbroid rocks, the most interesting are those of the olivine-bearing group. The principal type is an olivine norite, composed essentially of highly calcic plagioclase and olivine, with considerable hypersthene and hornblende. Through marked variations in the proportion of these minerals and the local appearance of augite in place of hypersthene, the olivine norite grades to troctolite (allivalite), olivine gabbro (eucrite), hornblende-olvine gabbro, and rarely anorthosite. However, the changes in the relative proportions of the mafic minerals are so common and local that the restricted terms are applicable only to particular hand specimens or outcrops, rather than to mappable units. All the rocks are dominantly feldspathic with over 60 per cent of plagioclase. The average grain size is about 1 mm., but this feature varies as frequently and erratically as the mineral content. The usual range is from less than 0.5 mm. to 3 or 4 mm., although in some small areas the grain size is as coarse as 20 mm.

In view of the calcic composition of their plagioclase, the olivine gabbros and troctolites fall within the definitions of the terms eucrite and allivalite, respectively.[7] The latter terms have not been used throughout the present paper, because they do not take into account the importance of hypersthene relative to augite and further because they have ordinarily been applied to rocks higher in mafic minerals and more definitely ultramafic in their affinities than those under consideration.

The two outstanding features of the olivine rocks are the extremely calcic composition of the plagioclase and the mosaic texture which this mineral commonly displays. The plagioclase is in the anorthite range, the average composition being An_{93}. It has an allotriomorphic equigranular texture of the type to which the writer has applied the term "mosaic".[8] In thin section the anorthite grains are anhedral, nearly equant, and bounded by straight or gently curving lines, which are independent of the crystal structure and which tend to give them polygonal shapes (Pl. 1, fig. 1). Ordinarily only the plagioclase shows the mosaic

[7] A. Harker: *The geology of the small isles of Inverness-shire*, Geol. Surv. Scotland, Mem. (1908) p. 68-101.

[8] F. S. Miller: *Anorthite from California*, Am. Mineral., vol. 20 (1935) p. 139-146.

texture, but locally olivine, augite, and in certain cases even hornblende also fit into the mosaic pattern.

In addition to the mosaic texture, the calcic plagioclase of the olivine rocks is characterized by lack of zoning and inclusions, uniformity of composition and optical properties, and only a limited development of twinning. It is significant that when even slight zoning is present, the average composition of the plagioclase drops to An_{87}–An_{90} and the mosaic texture is less well developed, the plagioclase then tending to form thick tablets with more or less definite crystal boundaries. This is true in rocks in which a considerable quantity of hornblende is present. The mosaic texture is best displayed in the rocks with the least hornblende and a minimum of zoning. Thus there is a definite relation between the uniform, unzoned, extremely calcic plagioclase and this particular texture.

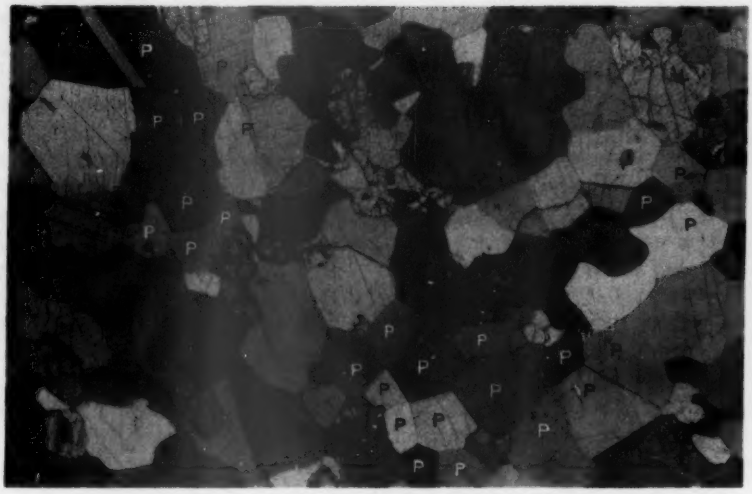
The amount of olivine in these rocks ranges from 7 to 20 per cent. The optical properties show it to be a chrysolite with 25 to 31 per cent iron orthosilicate.[9] The olivine is ordinarily separated from adjacent plagioclase by a characteristic reaction rim consisting of a narrow inner band of hypersthene next to the olivine and a wider outer band of pale-green hornblende intergrown with deep-green vermicular spinel. In fresh rocks most of the olivine is very slightly altered. Where extensively altered, the alteration is of a distinctive type which yields an intergrowth of graphic iron ore, iddingsite, and anthophyllite.

The typical hornblende of the olivine rocks has a compact habit and a very pale-green color, and the hypersthene and augite also appear unusually pale in thin section. This indicates a low Fe/Mg ratio in all three of these minerals. The optical properties of the hypersthene correspond to a composition range of 21 to 25 per cent iron metasilicate. The hypersthene, augite, and hornblende all tend to be either poikilitic or interstitial to the plagioclase and olivine, rather than prismatic in habit. In addition to the pale-green hornblende, some of the olivine rocks contain coarse-grained, poikilitic, green-brown hornblende. Those in which the alteration of olivine shows the influence of late reaction processes also contain minor pale fibrous hornblende and a fine-grained blue-green hornblende frequently associated with it. The olivine rocks are unusually low in iron ore, carrying only 0.1 to 0.3 per cent, except some of the olivine gabbros which have up to 2 or 3 per cent.

NORITES AND HYPERSTHENE GABBROS

The most widespread hornblende-poor rock is a typical norite. This is a medium-grained plutonic rock composed essentially of plagioclase and hypersthene with minor augite. Through a local increase in augite

[9] A. N. Winchell: *Elements of optical mineralogy*, Pt. 2, 3rd ed., New York (1933) p. 191.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relative to hypersthene, the rock grades to hypersthene gabbro without other change in appearance or composition. Since no fixed ratio of hypersthene to augite has been generally accepted as a requisite of the term norite, it does not seem necessary to draw a fine line between the two types. In the present case, rocks with hypersthene definitely predominant over augite in a ratio greater than 2:1 are termed norites. On this basis about one-fourth of the rocks of the group would be excluded from the norites. The norites are notably more uniform in mineral content and grain size than the olivine rocks. The average grain size is 1 to 2 mm., and the principal deviation from this arises through the appearance of a matrix of abundant, much smaller grains (0.05 to 0.4 mm.) in the midst of grains of the ordinary size. This produces seriate and porphyritic textures in some of the norite bodies. Yet on the whole the norites tend to maintain a given textural character over a much greater area than do the olivine rocks. The rocks with a porphyritic texture tend to be richer in augite than the other norites and so include a considerable proportion of the hypersthene gabbros.

The plagioclase of the typical norites is a labradorite with an average composition of An_{60}–An_{65}. It does not display the mosaic texture but occurs in well-developed tabular grains, ordinarily showing moderate zoning and sometimes a subparallel arrangement. An unusual feature is the occasional appearance of a core of bytownite-anorthite within some of the larger labradorite grains. The core may have a regular polygonal shape conformable to the crystal structure or consist merely of irregular patches or shreds or even a hollow shell (Pl. 1, figs. 2 and 3). The core material is in the composition range An_{80}–An_{92}, but chiefly An_{87}-An_{90}. Even the inner zones of the enclosing plagioclase have an anorthite content of only An_{60} to An_{70}. Thus there is a sharp composition break between the calcic cores and the adjacent labradorite, which represents a difference of more than 20 per cent in anorthite content.

The hypersthene and augite of the norites show a much stronger tendency to attain their characteristic prismatic form than those of the olivine rocks. The hypersthene has definitely higher refractive indices, stronger pleochroism, and a higher Fe/Mg ratio (28–32 per cent iron metasilicate) than the hypersthene of the olivine rocks.

Most of the hornblende of the norites is a compact deep-green hornblende derived largely from a reaction with the pyroxene. All stages are found in the centripetal replacement of hypersthene and augite by this reaction hornblende up to a point where only small remnants of pyroxene are enclosed poikilitically in large hornblende individuals. Ordinarily the replacement of augite by hornblende is further advanced

than that of hypersthene. Coarse, poikilitic, green-brown hornblende, apparently identical with that found in some of the olivine rocks, is also present in a number of norite specimens. Moreover, in three or four specimens, the plagioclase poikilitically included within the large green-brown hornblendes is anorthite, although the main plagioclase of the rock outside of the hornblendes is the usual labradorite. The inclusions are not merely tiny rounded blebs, but well-formed subtabular grains.

The norites contain more iron ore than the olivine rocks. The average is about 1 per cent by volume, but the porphyritic and finer-grained phases carry as much as 3 per cent, occasionally 5 per cent. In these ore-rich rocks augite is more prominent in comparison with hypersthene than in the typical norites, just as it is in those olivine rocks carrying appreciable iron ore.

Most of the norites contain neither quartz nor biotite, but a few have a very little biotite associated with the iron ore, and where there is a trace of biotite, there is often a trace of interstital quartz. When the amount of either of these minerals approaches 1 per cent, the rock is grouped with the quartz-biotite-bearing norites, for even this small amount is found to be indicative of the more sodic composition of the plagioclase. Similarly the typical norites are practically free from hornblende, although most of them have a small quantity of reaction hornblende adjacent to the pyroxene. Where only remnants of pyroxene remain in the midst of predominant hornblende, the rocks are classed with the hornblende gabbros. In such rocks there is again a tendency for the plagioclase to be slightly more sodic than in the typical hornblende-free norites. Those norites which contain considerable hornblende or traces of quartz and biotite, without sufficient to set them apart from the rest of the norites, tend to have a plagioclase in the composition range An_{55}–An_{60}, rather than in the higher range An_{60}–An_{65}.

QUARTZ-BIOTITE NORITES

The norites containing appreciable quantities of quartz and biotite are worthy of special mention because of their more sodic plagioclase. Quartz and biotite both begin to appear in rocks with a plagioclase more sodic than An_{56}–An_{58}. Norites with no distinctive mineralogical or textural features, other than the presence of 0.5 to 8 per cent each of quartz and biotite, have a sodic labradorite plagioclase with an average composition in the range An_{45}–An_{60}, usually An_{50}–An_{56}. In the same rocks the calcic cores in the plagioclase are more numerous and better developed than in the other norites. The absence of hornblende in as great a proportion of the quartz-biotite rocks as of the typical norites shows that these characteristics are not dependent upon the presence of hornblende. The combination of the more sodic plagioclase and the

appearance of quartz and biotite marks these rocks as one step farther away from the olivine rocks than the ordinary norites, and this is confirmed by the chemical analyses presented.

HORNBLENDE GABBROS

General Statement.—The hornblende gabbros as a group are the most complex of the San Marcos rocks. They are all composed essentially of plagioclase and hornblende, with only minor remnants of pyroxene, but the plagioclases cover such a wide range of composition and the hornblendes such a wide range of habit and color as to produce a confusing assemblage of rock types. Although the rocks are predominantly feldspathic, with 50 to 65 per cent of plagioclase, the amount of hornblende ranges from 20 per cent up to 50 per cent and so is frequently greater than the total of the mafic minerals in the hornblende-poor rocks. The tendency of the hornblende to resist weathering and to stand out prominently on weathered surfaces often gives a misleading impression of the amount present in the rocks.

The hornblende gabbros as a whole contain notably more iron ore than the other gabbroid rocks. The average is 2 to 4 per cent. Moreover, pyrite is prominent in a number of specimens in addition to the customary magnetite and ilmenite. The pyrite tends to be limited to the rocks which are richest in hornblende, especially those with abundant pale fibrous and blue-green hornblende.

Microscopic study shows that the hornblende gabbros fall into three fairly well defined divisions, which correspond respectively to the olivine rocks, norites, and quartz-biotite norites.

Calcic Hornblende Gabbros.—One group of hornblende gabbros is characterized by a highly calcic plagioclase, the composition of which does not regularly reach the An_{93} of the olivine rocks but ranges from An_{85} to An_{93}. The mosaic texture is present only locally; for the most part the plagioclase has less regular boundaries, or forms tabular grains which are much thicker in relation to their length than those of the norites. It also shows moderate zoning. Frequently the position of former olivine grains is marked by remnants of their alteration products —graphic magnetite, iddingsite, anthophyllite. Less common are remnants of reaction rims—vermicular spinel or original pale-green hornblende. The fact that these rocks are associated in the field with the olivine rocks further emphasizes the obvious relation between the two groups.

As in the other hornblende gabbros, the hornblende of the calcic rocks includes several different types. There are remnants of the pale-green hornblende of the olivine rocks. The coarse poikilitic green-brown hornblende is common and often abundant; some rocks are composed solely

of calcic plagioclase and this hornblende with no evidence of earlier pyroxene or olivine. At times the green-brown hornblende makes up as much as 50 per cent of the rock, although in the hornblende-poor rocks the maximum total for pyroxene and hornblende combined is about 35 per cent. In rocks where replacement of pyroxene and olivine by hornblende has reached an advanced stage, as is usual in the calcic hornblende gabbros, the pale fibrous and blue-green hornblendes tend to be prominent. Some specimens are made up entirely of calcic plagioclase and interstitial aggregates of these hornblendes, with a little associated chlorite. Since the pale fibrous and blue-green hornblendes attack even the other hornblendes and extend out into the plagioclase along fractures, they are the latest of the hornblendes.

Noritic Hornblende Gabbros.—Other hornblende gabbros stand in the same relation to the norites as the calcic hornblende gabbros do to the olivine rocks. With increasing amounts of hornblende surrounding pyroxene remnants, there is a complete gradation from norites to hornblende gabbros with no trace of former pyroxene. The plagioclase has the same range of composition as that of the norites (An_{55}–An_{65}) but the average is slightly more sodic (An_{57}). It also has the same characteristic tabular habit, zoning, and occasional calcic cores. The principal type of hornblende is the deep green reaction hornblende prevalent in the norites. Where the destruction of the pyroxene is complete, however, the pale fibrous and blue-green hornblendes are abundant. Coarse green-brown hornblende is also prominent in some of these rocks, but there is little evidence of its development at the expense of pyroxene to the same extent as is true of the other hornblendes. Hornblende-rich rocks corresponding to the porphyritic phases of the norites and hypersthene gabbros are also found.

Quartz-biotite-hornblende Gabbros.—Like the ordinary norites some of the hornblende gabbros contain small amounts of quartz and biotite. These rocks have a more sodic plagioclase than the quartz-biotite norites; the usual composition range is An_{45}–An_{56}. In this plagioclase calcic cores are slightly more abundant than in that of any of the other gabbros. With increasing quartz and biotite (compared to hornblende) and decreasing anorthite content in the plagioclase, the quartz-biotite-hornblende gabbros grade into the Green Valley tonalite. There is no sharp line of demarcation between the two types as regards composition, appearance, or field occurrence.

GREEN VALLEY TONALITE

It is beyond the scope of this paper to describe the Green Valley tonalite, but the features of this rock which set it apart from the San Marcos rocks and which at the same time fix the boundary between the gabbros

and the more felsic batholithic rocks are as follows: a lighter average color; a more sodic average composition of the plagioclase (An_{40}–An_{45}), with calcic cores more common than in the norites; quartz in excess of 10 per cent; biotite of equal importance with hornblende (about 8 to 15 per cent each); pyroxene definitely subordinate to both biotite and hornblende, less than 5 per cent occurring as remnants in these minerals; potash feldspar up to 5 per cent.

DIKE ROCKS

The well-known gem-bearing pegmatites of Pala and other neighboring localities occur to a large extent within bodies of the San Marcos rocks. Since these dikes have been described before by several writers[10] and seem to be related genetically to the granites and granodiorites of the batholith, rather than to the gabbros, they need not be discussed here. The numerous small aplite dikes which cut the gabbros are closely related in origin to the pegmatites, and presumably to the granodiorites.

The only dike rocks requiring discussion are the locally abundant lamprophyres in the gabbros. The dikes range from a fraction of an inch up to about a foot in thickness and are traceable for a few yards. The dike rock is very dark in color and for the most part fine grained. Since these dikes regularly belong to the same type of gabbro as the rock in which they occur, they appear to be strictly local in origin.

The lamprophyric dikes are most numerous in the olivine rocks and especially in the calcic hornblende gabbros associated with them. In these rocks the dikes are composed essentially of calcic plagioclase (An_{88}–An_{93}), green-brown hornblende, and iron ore (Table 1, No. 18). Dikes of this composition might be called anorthite spessartites or simply hornblende-anorthite dikes. Much of the anorthite shows a fine-scale mosaic texture with an average grain size of less than 0.2 millimeter. Part of the hornblende fits into this mosaic, but other larger hornblende grains, up to several millimeters long, are prismatic in habit. The dike rocks contain no olivine and only a very small amount of pyroxene, but they are unusually high in iron ore, chiefly magnetite.

Similar dikes occur to a lesser extent in the norites and in the hornblende gabbros derived from them. These dikes have a plagioclase with the same tabular habit and labradorite composition as that of the enclosing norite rocks. In the hornblende-free norites the dike rock is often

[10] G. A. Waring: *The pegmatite veins of Pala, San Diego County, California*, Am. Geol., vol. 35 (1905) p. 356-369.

G. F. Kunz: *Gems, jewelers' materials, and ornamental stones of California*, Calif. St. Min. Bur., Bull. 37 (1905) 171 pages.

F. J. H. Merrill: *Geology and mineral resources of San Diego and Imperial Counties, California*, Calif. St. Min. Bur., St. Mineral. Rept., pt. 5 (1913-1914) p. 636-732.

W. T. Schaller: *The genesis of lithium pegmatites*, Am. Jour. Sci., 5th ser., vol. 10 (1925) p. 269-279.

a fine-grained porphyritic hypersthene gabbro with very little hornblende. Such a dike is lower in dark minerals, for its pyroxene is not nearly so abundant as the hornblende of dikes in the hornblende-rich rocks. There is quartz in the one specimen from a dike in a quartz-bearing norite.

DISTRIBUTION OF THE GABBRO TYPES

The regional distribution of the different gabbro types and the field relations of one type to another are extremely complex, because each gabbro body commonly includes a wide variety of types and because at least locally almost complete gradation between the major types is found. The principal exception to this gradation is the gap between the olivine rocks and the norites. No rocks illustrating the appearance of olivine in small amounts are among the thin sections. The only rocks intermediate between the two groups are a few norites with calcic plagioclase in a mosaic texture and a single olivine rock with labradorite feldspar in a typical norite texture. No definite contacts were observed between the olivine rocks and the norites. The contacts between the other types are in large part gradational, although locally there is often a sharp demarcation between hornblende-rich and hornblende-poor rocks.

As an example of the complexity of distribution of the types within a single gabbro body, the Pala Mountain body may be cited. The northwest slope of this mountain up to the end of the summit ridge is made up predominantly of olivine rocks which vary from troctolite to olivine norite and olivine gabbro, together with irregular patches of calcic hornblende gabbro. On the westernmost spur of the mountain there is a large area of ordinary norite. On the subordinate ridges which extend down the southwest flank, both olivine rocks and norite occur in close association, and the same is true along the central portion of the summit ridge. At the south end of the mountain, norite and noritic hornblende gabbro are predominant, and there is a considerable mass of finer-grained porphyritic hypersthene gabbro. The northeast slopes are closed to investigation by the impenetrability of the brush, but along the foot of these slopes the rocks are quartz-biotite norite and the corresponding quartz-biotite hornblende gabbro. In short, all the principal varieties of the San Marcos rocks, both hornblende-poor and hornblende-rich, are found in this one body with an area of 4½ square miles. This is admittedly an extreme case. Yet the Monserate Mountain body and others show nearly as wide a range of types.

The main north-south ridge of the San Marcos Mountains constitutes the largest uniform mass of a single type of gabbro within the quadrangle. The rock is a typical norite, which is unusually constant in min-

eral composition and, except for a few local areas, surprisingly free from hornblende. The plagioclase falls consistently in the composition range An_{60}–An_{65}. Yet throughout the remainder of the same gabbro body the rocks present the usual variety of types. Down the ridges at the northwest end of the mountains in the vicinity of Vista Grande, the norite carries more hornblende and a more sodic plagioclase (An_{55}). Here there is a small area of nodular gabbro, and in this rock quartz and biotite, as well as hornblende, are present. At the south end of the mountains is the exceptional norite which contains olivine in the midst of tabular labradorite. The low range of hills half a mile west of the main peak of the San Marcos Mountains is made up predominantly of hornblende-rich rocks, which include a large area of calcic hornblende gabbro. In the lowlands between the west base of these hills and Vista, various types occur from norite to hornblende gabbro and quartz-biotite-hornblende gabbro. The plagioclase of these rocks ranges from An_{57} to An_{45} near the adjoining Green Valley tonalite. This description is much simplified and generalized.

In spite of the intricate distribution of the gabbro types, the following general tendencies may be pointed out: The concentration of the olivine rocks and calcic hornblende gabbros in the northeastern section of the quadrangle, the prominent development of the hornblende-rich rocks along the east side of the quadrangle from Pechanga to Valley Center, and the conspicuous development of the porphyritic rocks and quartz-biotite-bearing rocks from Vista to Escondido and south to the boundary of the quadrangle, in the same area in which the Green Valley tonalite is prominent.

The following is an estimate of the approximate area occupied by each of the principal gabbro types, considering 60 square miles as the total area of San Marcos rocks within the quadrangle.

Hornblende-poor rocks	Square miles	*Hornblende-rich rocks*	Square miles
Olivine rocks	8	Calcic hornblende gabbros	6
Norites and hypersthene gabbros	19	Noritic hornblende gabbros	18
Quartz-biotite norites	3	Quartz-biotite-hornblende gabbros	6
	30		30

CHEMISTRY AND MINERALOGY

Chemical analyses by F. A. Gonyer of rocks representing the main gabbro types are given in Table 2. Beneath them are shown norms calculated according to the standard C. I. P. W. procedure and modes determined by the Rosiwal method on thin sections of each analyzed rock. The modes are here expressed in weight percentages for direct comparison with the norms.

TABLE 2.—*Chemical analyses, norms, modes*

	3	20	7	13	15	22	23	25	A
Analyses									
SiO_2	42.86	48.16	52.12	47.22	56.52	58.68	63.70	64.86	49.50
TiO_2	.18	.76	.33	1.60	1.06	.68	.57	.55	.60
Al_2O_3	24.94	19.66	20.88	18.18	17.44	17.32	15.32	17.46	21.40
Fe_2O_3	2.13	1.53	.34	6.14	1.70	1.36	1.58	1.80	1.70
FeO	6.14	5.06	6.52	7.80	6.44	5.06	4.25	2.80	6.00
MnO	.06	.14	.09	.18	.06	.08	.09	.30	.10
MgO	9.28	7.59	6.90	4.93	4.65	4.46	3.18	1.36	6.90
CaO	13.08	13.46	10.14	10.46	7.92	7.52	5.68	5.26	10.80
Na_2O	.76	1.88	2.40	2.74	3.30	2.96	3.00	3.38	2.10
K_2O	.09	.31	.25	.13	.76	1.20	1.92	1.64	.37
H_2O	.78	1.48	.22	.46	.32	.67	.51	.50	.66
P_2O_5	.....	.11	.....	.....	.16	.06	.06	.05	.04
S	.03	.08	.....	.16	.01	.02	.....	.....	.04
	100.33	100.22	100.19	100.00	100.34	100.07	99.86	99.96	100.21
Norms									
Q	.....	.....	1.26	1.38	8.82	12.30	20.88	24.36	.....
Or	.56	1.67	1.11	.56	4.45	7.23	11.12	9.45	2.22
Ab	6.29	16.24	20.44	23.06	27.77	25.15	25.15	28.82	17.82
An	64.22	44.20	45.59	36.97	30.58	30.30	22.80	26.13	47.82
C	.....	.....	.....	.....	.....	.....	.....	.61	.....
Di	.68	17.39	3.76	12.17	6.37	5.66	4.54	.....	4.73
Hy	4.09	9.18	26.53	13.12	17.06	15.34	11.35	6.17	22.39
Ol	20.09	6.01	.....	.....	.....	.....	.....	.....	1.59
Mt	3.02	2.09	.46	8.82	2.55	2.09	2.32	2.55	2.55
Il	.46	1.52	.61	3.04	2.13	1.37	1.06	1.06	1.22
Py	.....	.12	.....	.24	.....	.....	.....	.....	.12
Ap	.....	.34	.....	.....	.34	.....	.....	.....	.....
	99.41	98.76	99.76	99.36	100.07	99.44	99.22	99.15	100.46
% An in Normal plag.	91	73	69	62	52	55	48	48	73
Modal plag.	93	c57	64	59	48	c52	c42	41	.....
Modes									
Plagioclase	67	47	65	61	59	55	48	49	.....
K Feldspar	.....	.....	.....	.....	.....	.....	.....	4	.....
Quartz	.....	.....	.....	1	8	16	24	25	.....
Biotite	.....	.....	.....	1	6	8	13	15	.....
Hornblende	11	46	0.9	13	1.4	10	10	7	.....
Pyroxene	2	6	33	16	22	9	4	.....	.....
Olivine	18	.....	.....	.....	.....	.....	.....	.....	.....
Fe Ore	0.2	1.9	1.7	8	4	3	0.9	.....	.....
Spinel	2	.....	.....	.....	.....	.....	.....	.....	.....
Iddingsite	0.2	.....	.....	.....	.....	.....	.....	.....	.....
Chlorite	.....	0.1	.....	.....	.....	.....	.....	.....	.....
	100.4	101.0	100.6	100	100.4	101	99.9	100	.....

3. Olivine norite (with hornblende); northwest end of summit ridge of Pala Mountain; elevation, 2050 feet.

20. Noritic hornblende gabbro; crest of slope overlooking Santa Margarita River from south, 1½ miles west of Fallbrook.

7. Typical norite (hornblende-free); summit of San Marcos Mountain at B. M. 1664 feet, 3 miles east of Vista.

13. Norite with hornblende, quartz, biotite; quarry south of Escondido Creek, 3 miles west-southwest of Escondido.

15. Quartz-biotite norite; high level valley in northeast part of Monserate Mountain body; elevation, 1100 feet; 2½ miles west-northwest of Pala.

In Figure 3 the oxides of the analyses are plotted against silica in the usual form of variation diagram. Plutonic rocks such as the San Marcos gabbros would hardly be expected to conform to so simple a type of

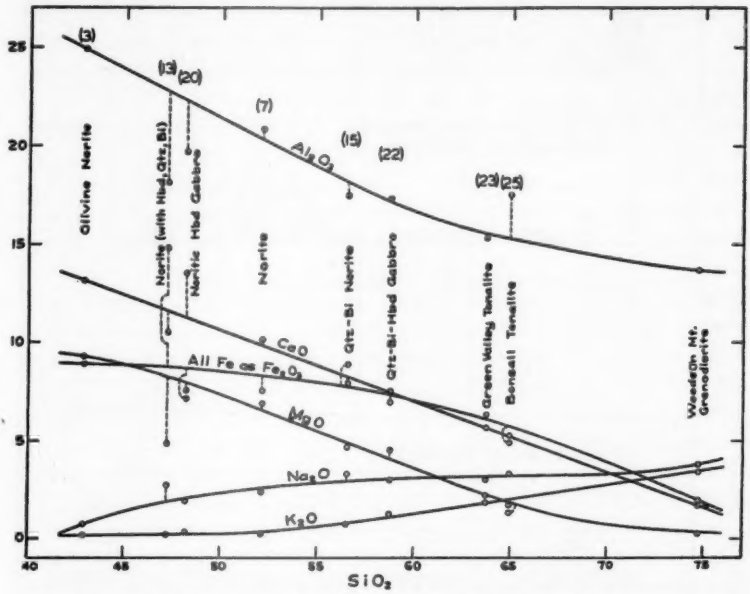

FIGURE 3.—*Variation diagram of the San Marcos Gabbros and related rocks*

diagram as the rocks of volcanic provinces. Crystal accumulation, movement of residual liquids, or assimilation of foreign material might have caused them to depart from a direct liquid line of descent. Yet on the whole the plotted oxides fit satisfactorily on even curves. The two aberrant analyses (Nos. 13 and 20) have not been given equal weight in drawing the curves, because they come from rocks which contain an abnormal amount of hornblende and which appear on the diagram much nearer the low-silica end than is warranted by the composition of their modal plagioclase and other critical features. This probably results from

22. Quartz-biotite-hornblende gabbro, intermediate between San Marcos rocks and Green Valley tonalite; road cut 1½ miles west of Vista on road to Oceanside.

23. Green Valley tonalite; road cut 2 miles west-northwest of southeast corner of San Luis Rey quadrangle.

25. Bonsall tonalite; 2 miles northeast of San Luis Rey Mission. Analysis quoted from Hurlbut, *op cit.*, p. 613.

A. Average chemical composition of San Marcos gabbro computed on basis of relative abundance of rocks corresponding to analyses 3, 20, 7, 13, 15 and 22.

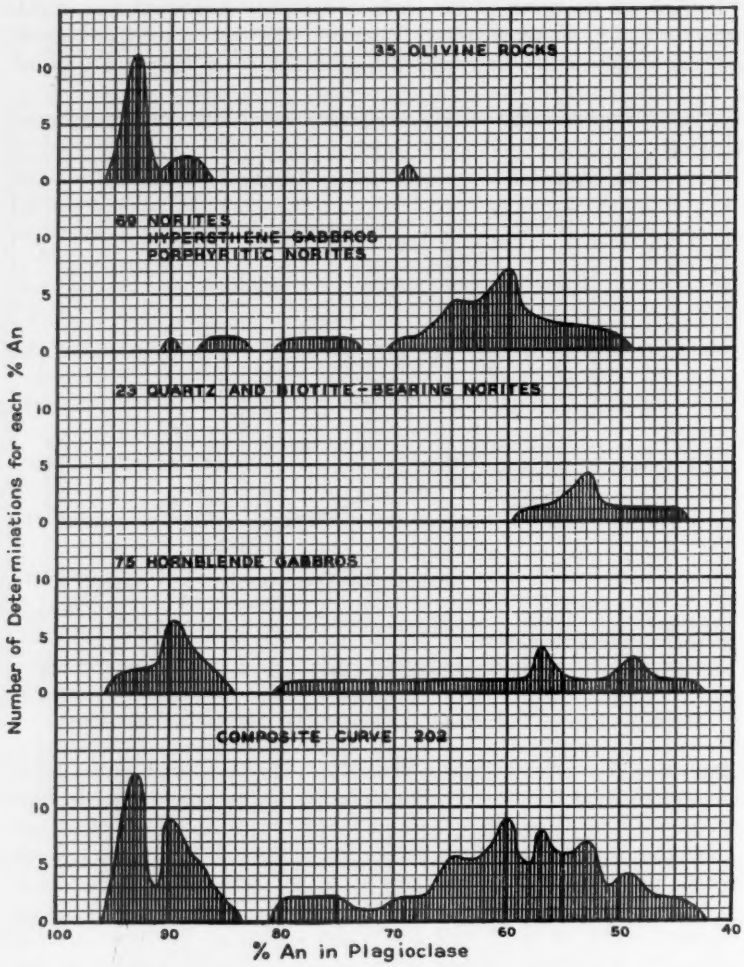

FIGURE 4.—*Distribution of the plagioclase by composition*

a silica deficiency in the hornblende. The unusual character of the hornblende-rich rock No. 20 is also evidenced by the discrepancy in composition between its normative plagioclase and its definitely more sodic modal plagioclase (Table 2). In addition to these complications arising in connection with the hornblende, the erratic distribution of late-formed

iron ore provides a disturbing factor in some of the rocks. This is illustrated by the unusually high iron ore content of No. 13. It is noteworthy that the silica content of the olivine norite (No. 3) is so low that it causes an extension of the variation diagram for this series of feldspathic rocks into the low silica range ordinarily occupied only by the ultramafic rocks. The uniform straight-line descent of the lime points to an interesting direct ratio between lime and silica.

During the laboratory study of the San Marcos rocks the composition of the plagioclase in more than 200 specimens was accurately determined by the immersion method. The results of these determinations are plotted in Figure 4, in such a way as to make clear the relative abundance of plagioclase of different compositions and also the relation of plagioclase composition to rock type. The significant variations in plagioclase composition may be summed up as follows:

1. The anorthite plagioclase of the olivine rocks falls consistently between An_{92} and An_{95}. The few specimens with a plagioclase between An_{87} and An_{90} show moderate zoning and less well-developed mosaic textures. There is a greater concentration of plagioclase at An_{93}-An_{94} than at any other place in the diagram. Moreover, the determinations should be the most accurate for this plagioclase, because it has a minimum of zoning. Therefore there is a greater abundance of rock with plagioclase of exactly this composition than of any other, a fact which is significant in view of the erratic variations in the composition of the plagioclase in many of the calcic rocks described in the literature.

2. The ordinary norites and hypersthene gabbros are most abundant in the range An_{60}-An_{65}. Most of those in the range An_{55}-An_{60} carry considerable hornblende or traces of quartz and biotite. The porphyritic norites are distributed through the range An_{50}-An_{65} but are especially prominent between An_{50} and An_{55}. The few calcic norites with plagioclases up to An_{90} have textures which approach the mosaic.

3. The quartz- and biotite-bearing norites are concentrated in the range An_{50}-An_{56}.

4. The curve for the hornblende gabbros has three separate peaks, which correspond respectively to the three main hornblende-poor types, but each peak is displaced in the sodic direction 3 to 4 per cent from the peak of the hornblende-poor equivalent.

Parallel to the changes in the composition of the plagioclase, systematic variations would be expected in the composition of those mafic minerals which belong to isomorphous series. Of such minerals hypersthene is the only one which occurs widely throughout the gabbroid rocks and which at the same time has a composition simple enough to be

reflected accurately by its optical properties. The increase in the Fe/Mg ratio of the hypersthene in direct relation to the increasing Ab/An ratio of the plagioclase is shown by the following table:

	4 Rocks with An_{59}–An_{55}	*3 Rocks with* An_{55}–An_{51}	*3 Rocks with* An_{47}–An_{33}
α	1.683–1.689	1.696–1.698	1.703–1.706
β	1.691–1.697	1.705–1.709	1.714–1.716
γ	1.696–1.702	1.710–1.713	1.719–1.721
γ–α	0.012–0.013	0.014–0.015	0.015–0.016
% $FeSiO_3$	21–25	28–30	33–34

The slightly higher refractive indices of the augite in the more sodic rocks probably indicate a similar increase in the Fe/Mg ratio of this mineral. But the differences are small and are not susceptible to simple interpretation because of the complex composition of the augites. The range of the optical properties of the augites is as follows: $\alpha = 1.686$-1.694, $\beta = 1.691$-1.700, $\gamma = 1.710$-1.719, $\gamma-\alpha = 0.023$-0.026, $+2V = 55°$-$62°$, $r > v$ perceptible, $Z \wedge c$ $39°$-$44°$.

Olivine and biotite are each limited in occurrence to a single group of rocks, and the two groups come at opposite ends of the plagioclase series. The optical properties of the olivine are: $\alpha = 1.682$-1.693, $\beta = 1.704$-1.716, $\gamma = 1.720$-1.735, $\gamma-\alpha = 0.038$-0.044, $-2V = 80°$-$86°$, $r > v$ distinct. These indicate a range of composition from 25 to 31 per cent Fe_2SiO_4. The properties of the biotite show a considerable iron content: $\alpha = 1.595$–1.609, $\beta = 1.655$–1.672, $\gamma = 1.660$–1.675, $\gamma-\alpha = 0.062$–0.066, $-2V = 3°$-$10°$, $r < v$ weak.

The variation in the optical properties of the hornblende is not so great as might be expected from the pronounced differences in its habit, color, and occurrence. Five types of hornblende have been mentioned in the rock descriptions, but the range of optical properties within a given hornblende type is sufficient to overlap to a large extent those of the other types. The following features may be indicative of differences in the Fe/Mg ratio among the hornblendes: (1) The pale-green hornblende of the olivine rocks has refractive indices slightly lower than any of the other hornblendes and is the only optically positive hornblende. (2) The deep-green reaction hornblende of the norites has the widest range of refractive indices of any single hornblende type, but the higher indices are found only in the quartz- and biotite-bearing rocks, *i.e.*, the more sodic rocks. The optical properties of the various types of hornblende will be presented in another paper in connection with a general discussion of the types and their probable manner of formation.

THEORETICAL CONSIDERATIONS

GENERAL CONSIDERATIONS

A consideration of the field data and the textural, mineralogical, and chemical features of the San Marcos rocks yields some interesting conclusions with regard to their mode of origin.

CALCIC MAGMA

The observed features of the olivine rocks seem to establish definitely the existence of a magma not merely calcic enough to precipitate anorthite, but actually with a total composition as calcic as the prevalent An_{93} plagioclase of those rocks.[11] The evidence for the existence of such a calcic liquid is as follows:

1. The consistency with which all the plagioclase of most of the olivine rocks falls at the precise composition An_{93}–An_{94}.
2. The absence of zoning and the uniformity in optical properties of this anorthite.
3. The primary nature of the mosaic texture.
4. The occurrence of fine-grained lamprophyric dikes composed solely of hornblende and anorthite.
5. The conspicuous absence of either rocks or dikes with a plagioclase in the composition range An_{80}–An_{85} to represent the liquid from which the An_{93}–An_{94} plagioclase would have been precipitated, if formed by fractional crystallization.

The critical feature of the mosaic texture is the lack of any relationship between the grain boundaries and the crystal structure. The mosaic pattern appears to have been produced by the mutual interference of growing anorthite crystals, which filled up all the available space, so crowding each other that none had opportunity to develop its usual tabular form. This type of crystallization would preclude the possibility of differentiation by the accumulation of crystals, or the escape of residual liquids, and would produce a rock of essentially the same composition as the local magma from which it crystallized.

The similarity of the mosaic texture to the saccharoidal texture of many aplite dikes is noteworthy, for in the aplites the texture is generally considered to denote a simultaneous crystallization such as that just described.[12] The apparent similarity of the mosaic pattern to the texture found in some crystalline limestones and quartzites does not reflect on the primary nature of the mosaic texture in the olivine rocks, since the latter rocks show no signs of metamorphism. The only post-magmatic changes of which there is any evidence in these rocks consist of

[11] N. L. Bowen: *The evolution of the igneous rocks* (1928) p. 143, Princeton.
[12] Albert Johannsen: *A descriptive petrography of the igneous rocks*, vol. 2, Chicago (1932) p. 92.

the alteration of olivine and the introduction of late hornblende and chlorite at the close of the magmatic period. Where the mosaic texture is best developed, the olivine is still fresh and the rock is free from evidence of even these slight changes.

The discovery of the fine-grained hornblende-anorthite dikes leaves no room for doubt concerning the existence of a calcic liquid. These dikes, associated only with the calcic rocks, must represent residual magma fractions locally expelled from those rocks. To produce the dikes it is only necessary to segregate from the crystallizing calcic hornblende gabbros a residual liquid slightly enriched in hornblende and iron ore and then to inject this into a neighboring portion of the mass already sufficiently consolidated to fracture. The composition of the plagioclase in individual dikes reaches as high as An_{93}, but in most cases it is only An_{88}–An_{90}. This is what would be expected if the dikes are segregations from a late stage of the crystallization.

In dealing with these calcic plagioclases it is important to have in mind how closely the liquidus and solidus curves of the plagioclase equilibrium diagram converge within the anorthite range. In some of the olivine rocks the plagioclase has a uniform composition as high as An_{95}. According to Bowen's curves [13] plagioclase of this composition would be in equilibrium with a liquid of the composition An_{85} and so must have been precipitated from a liquid at least that calcic. The prevalent An_{93}–An_{94} plagioclase would be in equilibrium with a liquid of the composition An_{80}–An_{85}. Since the necessity of accounting for a liquid as calcic as An_{85} is therefore inescapable, any objection to an An_{93} magma on the basis of the difficulty of explaining its origin by current theories of magmatic differentiation loses much of its weight.

ORIGIN OF MOSAIC TEXTURE

Although in these rocks the mosaic texture is characteristic only of the anorthite and appears to be dependent upon the presence of unzoned highly calcic plagioclase, it is also shown locally by augite and olivine, and in the hornblende-anorthite dikes by hornblende, in conjunction with the anorthite. The pattern must therefore have resulted from the conditions of crystallization, and not from any inherent weakness in the tendency of the anorthite to attain its crystal form. There is an inverse relationship between the mosaic texture and the distribution of hornblende (except in the dikes). The mosaic pattern is best developed in the rocks composed simply of olivine and anorthite—the troctolites—

[13] N. L. Bowen: *The melting phenomena of the plagioclase feldspars*, Am. Jour. Sci., 4th ser., vol. 35 (1913) p. 577-599.

and is not well shown in the calcic hornblende gabbros, nor in those parts of the olivine rocks where hornblende is prominent. Since hornblende requires water for its formation, and so may be taken to indicate the presence of mineralizers in the magma, the mosaic texture is most pronounced where mineralizers were last abundant.

It is likely that the contrast in texture between the troctolites and the calcic hornblende gabbros is attributable directly to the local differences in composition within the calcic magma, which produced the two rock types, particularly the differences in mineralizer content. The nature of the mosaic pattern requires that it be formed from a liquid of such composition that the components were able to crystallize simultaneously and completely without yielding a residual liquid capable of precipitating interstitial minerals. In the portions of the calcic magma which formed the troctolites, the simple composition of the magma and the absence of mineralizers would combine to favor simultaneous crystallization and to enable it to go to completion without the development of an interstitial liquid. In a magma of this character there are also several factors which would aid in the development of uniform unzoned anorthite: (1) The high temperature of crystallization at the anorthite end of the plagioclase series would favor rapid diffusion in the crystallizing anorthite. (2) The absence of mineralizers would eliminate a factor tending to lower this temperature of crystallization and at the same time would increase the viscosity of the magma. (3) The liquidus and solidus curves are so close together in this range that little reaction would be necessary to bring the plagioclase to equilibrium. The equant habit of the anorthite crystals might also assist in the development of the mosaic texture.

The calcic hornblende gabbros, on the other hand, represent portions of the calcic magma in which were concentrated the mineralizers and other constituents necessary to produce abundant hornblende. The crystallization of these fractions would yield an interstitial liquid which would tend to facilitate the formation of tabular plagioclase and to rework earlier olivine and pyroxene. The presence of mineralizers would tend to slow down the rate of diffusion within the solid plagioclase by lowering the temperature of crystallization and at the same time to speed up the rate of diffusion in the liquid by decreasing its viscosity. This may account for the slight zoning of the plagioclase in these rocks. The development of a mosaic or submosaic texture in the fine-grained hornblende-anorthite dikes shows that under the circumstances of crystallization of these dikes even hornblende and anorthite were able to crystallize simultaneously.

CALCIC CORES

It is difficult to account for the origin of the calcic cores found occasionally in the plagioclase grains of the norites and more frequently in those of the quartz-biotite norites and the Green Valley tonalite, in spite of an obvious source for the material in the calcic plagioclase of the olivine rocks. The difficulty lies in the pronounced composition break between the anorthite or calcic bytownite of the cores and the surrounding labradorite. The fact that many of the cores have irregular embayed shapes or have been reduced to mere hollow shells indicates that the calcic crystals of which they are the remnants were subjected to an active solution, rather than to a simple reworking by a more sodic liquid. These calcic crystals must have been far out of equilibrium with the liquid from which the encasing labradorite was deposited.

The obvious assumption that the calcic crystals settled or floated into a much less calcic liquid with which they were not in equilibrium is not satisfactory. If calcic crystals are to settle into a more sodic liquid, there must be above the sodic liquid a calcic liquid capable of precipitating the crystals. This would require liquid immiscibilty or very imperfect mixing. Besides, the calcic liquid would have a higher specific gravity than the sodic liquid beneath it, and if it had a low enough viscosity over a long enough time to allow crystals to settle, the calcic liquid itself should have settled bodily. The floating of the calcic crystals presents equal difficulties, for it is very doubtful whether calcic plagioclase would float in a tonalite magma, or in the felsic part of a gabbro magma.

The most reasonable explanation of the origin of the cores is that they are remnants of assimilated calcic plagioclase material which the norite and tonalite magmas derived from the calcic rocks during intrusion. Hurlbut's study of the inclusions in the Bonsall tonalite[14] has demonstrated the importance of such assimilation in the most widespread plutonic rock of the area. He showed that the inclusions and schlieren so widely disseminated throughout the Bonsall tonalite were derived from foreign rocks, largely the San Marcos gabbro. The abundance of these inclusions is surprising. They commonly make up 5 to 10 per cent of the rock by volume, and locally in small areas as much as 30 per cent. Some are present in practically every outcrop of the tonalite over its total area of 100 square miles within the quadrangle. Moreover, partially reworked inclusions of gabbro in Bonsall tonalite show typical calcic cores which are not ordinarily found in the tonalite. It appears that the calcic plagioclase is the last thing to be completely reworked

[14] C. S. Hurlbut: *Dark inclusions in a tonalite of southern California*, Am. Mineral., vol. 20 (1935) p. 609-630.

in the process of reaction of this tonalite magma upon inclusions of gabbro.

Similarly, the calcic cores of the norites and Green Valley tonalite may be remnants of calcic inclusions in the last stages of digestion by the respective magmas. It would be natural to expect the norite magma, which was nearer in composition to the calcic rocks, to be better able to rework inclusions of those rocks than the tonalite magma. Thus fragments of calcic rock engulfed in the norite magma would have disappeared almost entirely before the final crystallization of the rock, leaving only a few scattered remnants of calcic plagioclase. The same fragments in the Green Valley magma would leave more numerous calcic remnants, because of the less thorough reworking of the fragments by the cooler, more felsic magma. In the Bonsall tonalite the inclusions themselves are found, in various stages of digestion, but still with calcic cores occurring prominently.

The principal question with regard to this explanation of the origin of the cores is the matter of its quantitative sufficiency. In view of the thoroughness with which reaction inclusions are disseminated through the Bonsall tonalite, the process of incomplete assimilation seems altogether sufficient to account for the scattered calcic cores in the San Marcos rocks. To apply the same process to the Green Valley tonalite, which shows cores in nearly every thin section, would strain the mechanism to the limit. Yet if it took place in depth and if there was mixing on a small scale when the reworking of the inclusions was already well advanced, such assimilation would be adequate to explain the cores even in the Green Valley rocks.

Since it is probable that the intrusion of the calcic magma slightly preceded that of the norite magma, the latter must have encountered a large amount of calcic material during its emplacement. The calcic rocks may have cooled very little before the intrusion of the norites, and, as they would crystallize at a higher temperature than the norites, they may still have been nearly as hot as the norite magma at the time of its intrusion. Inclusions of calcic material in the norite magma would thus require very little heating up previous to reworking, and the reaction process would be greatly facilitated. Under the same conditions any inclusions of the pre-batholithic quartzites and schists would be completely reworked.

To carry this reasoning a step further, the calcic magma itself might not have been thoroughly crystallized when the norite magma was intruded. The inclusions assimilated by the latter magma would then have merely been fragments of a crystal mesh of the partly solidified

calcic magma, or even separate crystals from a mesh which had been entirely disrupted and disaggregated. The only limit to this line of reasoning is that the calcic magma must have crystallized sufficiently to provide the calcic plagioclase grains and also presumably to have lost its mobility. In rocks of no greater difference in composition than the calcic rocks and the norites, the assimilation of even a large amount of calcic material would have comparatively little effect upon the composition of the norite magma.

Rittmann[15] has described almost identical calcic cores in a hornblende gabbro from the small islands of Kellang and Manipa near Ceram in the Dutch East Indies. Cores of bytownite-anorthite (An_{85}–An_{95}) are separated from surrounding borders of labradorite (An_{55}–An_{60}) by a sharp composition break, which he calls a "Zonierungsdiskordanz durch Korrosion verursacht". It is significant that the hornblende gabbro which carries these cores is closely associated with an olivine gabbro composed predominantly of bytownite-anorthite. Rittman suggests that after the crystallization of the cores, there was an addition of plagioclase aplite magma to the crystallizing hornblende gabbro; this so changed the composition of the latter that the bytownite-anorthite became no longer stable and was actively corroded. On further cooling the crystallization began again with a labradorite. However, in view of the slow mixing of two magmas by diffusion, it seems doubtful that such mixing would produce the observed composition break and corrosion of the cores.

MAGMATIC DIFFERENTIATION

In discussing the origin of the calcic cores, the intrusion of a norite magma was mentioned. Evidence for the existence of such a magma is the occurrence of fine-grained norites, porphyritic rocks with a fine-grained noritic groundmass, and fine-grained lamprophyric dikes having the same relations to the norites as the hornblende-anorthite dikes have to the olivine rocks. In addition, the norites are more consistently uniform in character and form larger uniform masses than any of the other gabbro rocks.

Moreover, all the rocks of the norite group, that is, the norites, quartz-biotite norites, and corresponding hornblende gabbros, are so closely related and grade into one another so completely that they obviously have been derived from a single parent magma by a systematic process of differentiation. The Green Valley tonalite, into which the quartz-biotite-hornblende gabbros pass without any break, may also be added to the assumed differentiation series. Within these rocks plagioclases

[15] A. Rittmann: *Gesteine von Kellang und Manipa*, Geol., pet., and pal. results of explorations in the island of Ceram (L. Rutten and W. Hotz) Amsterdam, 1st ser., no. 2 (1931) esp. p. 63-65.

of all compositions from An_{80} to An_{30} are represented in a practically continuous series, and the Fe/Mg ratio of the mafic minerals increases in the same sense as the Ab/An ratio of the plagioclase. Quartz and biotite appear at a definite point and increase in roughly parallel fashion, as the Ab/An ratio increases. In the tonalite a little potash feldspar comes in as well. These facts suggest that the rocks crystallized from a magma which was being progressively enriched in silica, soda, and potash through the abstraction of crystals rich in lime and magnesia. In spite of the apparent gap between the calcic rocks and the norites, which is wider than that between any of the rocks just mentioned, the calcic rocks are so closely associated with these rocks and have so many features in common with them (*see* p. 1399) that they must have come from the same ultimate source and presumably belong to the same differentiation series. It is probable that the series must also be extended at the felsic end to include the other tonalites and granodiorites of the batholith.

Yet any attempt to explain the relation between the calcic magma and the norite magma by deriving one of them from a parent magma similar in composition to the other meets with serious difficulties. To derive the calcic magma from the norite magma by fractional crystallization would require not only the melting of settled crystals, but repeated remeltings such as proposed by Vogt.[16] Plagioclase of the composition An_{95} would be in equilibrium with a liquid of the composition An_{85}, and plagioclase of the latter composition would in turn be in equilibrium with a liquid of the composition An_{62}, the composition of the plagioclase of the norites. Thus at least two remeltings would be necessary, even if it were possible to obtain each time a complete separation of the most calcic crystals, and this is unthinkable. By necessitating still further remelting the process becomes too complicated and the required adjustments too delicate to be regarded as practicable. Yet it is possible.

The alternative of deriving the norite rocks from a parent calcic magma by ordinary fractional crystallization is equally difficult. If it is necessary to extend the process to include the more felsic batholithic rocks, that makes it just so much more difficult. An enormous number of early-formed crystals would have to be removed in order to give a sufficient increase in the silica, soda, and potash content of the liquid. For instance, if the amount of potash in the analyzed olivine norite (Table 2, No. 3) is taken as the unit, the ratio of increase necessary to produce the norite (No. 7) is 2½ times, the quartz-biotite norite

[16] J. H. L. Vogt: *The physical chemistry of the magmatic differentiation of igneous rocks*, Videnak.-Selsk. Skr., I, Mat.-Naturv. Kl., Oslo, no. 15 (1924) p. 87.

(No. 15) 8 times, the Green Valley tonalite (No. 23) 19 times, and the Woodson Mountain granodiorite 34 times. This does not mean that the crystallization of 34 volumes of olivine norite would provide the potash for one volume of granodiorite, for the olivine rocks themselves and all the intervening rocks in the assumed differentiation series would also take up potash in the proportions indicated. The low potash content of the calcic rocks is an important obstacle to the derivation of the norites and more felsic rocks from the calcic magma by any simple scheme of differentiation. The soda and silica offer similar but quantitatively less serious difficulties. Yet the same difficulty remains in the case of deriving the more felsic rocks from the norite magma, even though the differentiation series is here apparently continuous from the norite at least as far as the Green Valley tonalite. The ratio of concentration of potash from the norite to the quartz-biotite norite is 3 times, to the Green Valley tonalite 7 times, and to the Woodson Mountain granodiorite 13 times.

The part played in the differentiation process by the assimilation of country rocks during intrusion is conjectural, but the respective chemical compositions of the batholithic and pre-batholithic rocks set definite limits to the importance of this factor. It may be assumed that inclusions of the schist and quartzite country rocks were highly silicic and contained soda and potash in roughly equal amounts, varying on the average from twice as much of the one to twice as much of the other. An examination of the analyses shows that assimilation could have played very little part in a differentiation of the norite from the olivine norite, for the addition of sufficient xenolithic material to account for the increase in soda would introduce from 5 to 10 times too much potash. At the felsic end of the differentiation series from the quartz-biotite norite to the granodiorite the soda remains relatively constant, and the situation is reversed. Here the addition of sufficient foreign material to effect the required increase in potash would bring in far too much soda. Also at this end of the series each successive intrusive phase would encounter fewer pre-batholithic rocks, while the assimilation of earlier batholithic rocks would have retarded rather than assisted the differentiation.

Thus assimilation cannot have been a major factor in the differentiation of the successive gabbroid and more felsic rocks, and yet, in view of the evidence of the abundant inclusions in the Bonsall tonalite and the calcic cores, it probably did have a definite minor effect. Although the low potash content of the norite sets a limit to the influence of assimilation as a factor in a differentiation of this rock from the olivine

norite, it is at the same time possible that the assimilation of potash-bearing material within this limit may have helped to provide the increase of potash in the norite which was difficult to account for by simple fractional crystallization of the calcic magma.

SUMMARY OF CONCLUSIONS

The conclusions from the foregoing considerations may be summarized as follows:

(1) All the San Marcos rocks were derived ultimately from the same parent magma. The rocks from the norite through the quartz-biotite norite and the quartz-biotite-hornblende gabbro to the closely related Green Valley tonalite represent a continuous differentiation series, in the production of which fractional crystallization appears to have been the dominant factor. Evidence of this is found in the regular increase of the Ab/An ratio of the plagioclase and the Fe/Mg ratio of the mafic minerals parallel to the increase in silica content throughout the series.

(2) The calcic cores in the plagioclase of these rocks are the last remnants of inclusions of calcic plagioclase material assimilated in depth during the early stages of intrusion. Yet it is improbable that assimilation had sufficient effect upon the composition of the magma to be more than a minor factor in the process of differentiation.

(3) The calcic gabbros (olivine rocks and calcic hornblende gabbros) crystallized directly from a magma as calcic as their predominant plagioclase An_{93}. This calcic magma may have been derived from the norite magma by some special differentiation process, such as the repeated remelting of settled crystals, or may itself have been the common parent magma of the norites and the more felsic rocks.

(4) All the gabbros were intruded almost contemporaneously, with the calcic phases perhaps a little in advance of the main norite magma. These magmas may have been caught in the midst of deep-seated differentiation at the time when the intrusion began, so that the intruded bodies were heterogeneous in composition from the beginning. The irregular distribution of hornblende, which divides the gabbros into hornblende-poor and hornblende-rich types, reflects an original erratic distribution of mineralizers in the intruded magma.

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TIMISKAMING SUB-PROVINCE

BY W. H. COLLINS
(With an introductory note by C. K. Leith)

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INTRODUCTORY NOTE: BY C. K. LEITH

In the summer of 1936, Dr. Collins proposed a book on the geologic history of "The Canadian Shield", under the joint authorship of spe-

cialists familiar with different parts of the region, and prepared an "experimental draft" of a table of contents. Following this proposal, it was arranged that the book should be prepared by W. H. Collins, Morley E. Wilson, T. L. Tanton, Andrew Leith, and C. K. Leith.

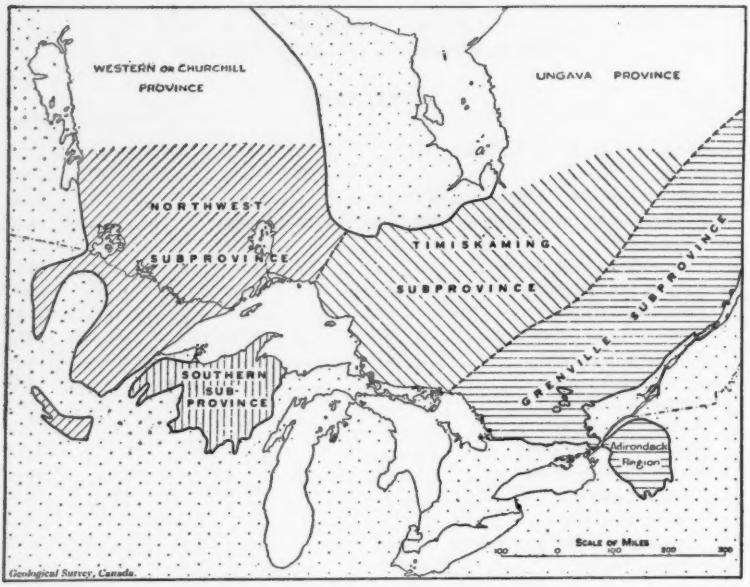

FIGURE 1.—*Sub-provinces of the Saint Lawrence geological province*

At the time of Dr. Collins' untimely death, he had prepared a preliminary draft of a chapter of this book on "The Timiskaming Sub-province", which presented such a good review of this part of the pre-Cambrian Shield as to warrant publication by itself, particularly in view of the probability that publication of the proposed book might be considerably postponed.

The manuscript has been edited only to the extent of eliminating cross references to other parts of the proposed book. The conclusions as to correlation have not been jointly reviewed by the co-authors, and are to be regarded as Dr. Collins' personal views. Had he lived, it is probable that in the proposed joint publication these conclusions might have been qualified to some extent by evidence from other parts of the Canadian Shield. Particularly do we have in mind the desirability of a more comprehensive statement of the problem of correlation of the Timis-

kaming and Sudbury series with the Knife Lake slates of the Lake Superior region. However, the manuscript as it stands is a valuable contribution from a geologist who probably knew, better than any other, this section of the pre-Cambrian Shield.

INTRODUCTION

The Timiskaming sub-province is chosen first among the major divisions of the Canadian Shield for an account of what is known about its geological history because systematic geological work was begun there earlier than in the neighboring sub-provinces south and west of Lake Superior. In consequence, the discoveries and writings of the first workers in this sub-province exerted considerable influence on geological thought in these other sub-provinces.

The position and extent of the Timiskaming sub-province and the nature of the geological barriers which separate it from each of the three neighboring sub-provinces are depicted in Figure 1. A geological map of the sub-province to illustrate this paper is impracticable. It would be prohibitively expensive and large. Readers are therefore referred to the best existing maps, the Lake Huron 8-mile map, No. 155A, and the Nottaway 8-mile map, No. 190A, both published by the Geological Survey of Canada and the Ontario Department of Mines, which will be mentioned here and there in this paper. The Lake Huron sheet alone will serve most map needs, for it represents a large area and nearly the entire geological succession found in the sub-province.

John Bigsby (1821), a surgeon of the British army stationed in Canada, made the earliest recorded geological observations of any part of Timiskaming sub-province while he was on a journey up the Great Lakes on other business. Continuing and systematic geological work was not begun, however, until after creation of the Geological Survey of Canada. When the first united parliament of Upper and Lower Canada (now Ontario and Quebec) met in 1841, a petition for a geological survey of the country was presented by the Natural History Society of Montreal and the Literary and Historical Society of Quebec. On motion of the Honorable G. B. Harrison on September 10, 1841, it was resolved "That a sum of money not exceeding one thousand five hundred pounds, sterling, be granted to Her Majesty to defray the probable expense in causing a Geological Survey of the Province of Canada." William Edmond Logan, a Canadian educated in Great Britain, was appointed provincial geologist in 1842, and he employed an assistant geologist, Alexander Murray, the next year.

Between 1845 and 1858, these two explored various parts of the country between the foot of Lake Timiskaming and Sault Ste. Marie, being attracted there by the discovery at Bruce Mines in 1846 of copper in the form of chalcopyrite within quartz veins. The results of these explorations are detailed in the annual Reports of Progress of the Geological Survey for that period and in other articles, a list of which, with summaries of their contents, is given by Van Hise and Leith in Bulletin 360 of the United States Geological Survey. This admirable bulletin should be used as a supplement for this and all other works on the pre-Cambrian Shield. The works of Logan and Murray are also summarized in the Geology of Canada 1863 and accompanying atlas. The result of chief interest was the discovery and stratigraphic description of the Huronian succession.

HURONIAN SUCCESSION OF LOGAN AND MURRAY

Logan and Murray discovered a succession of formations consisting in ascending order of:

	Feet
Gray quartzite	500
Chloritic slates	2000
White quartzite	1000
Lower slate conglomerate	1280
Limestone	300
Upper slate conglomerate	3000
Red quartzite	2300
Red jasper conglomerate	2150
White quartzite	2970
Yellow chert and limestone	200
White quartzite	400

These beds are folded up to angles of about 45 degrees. The country is now partly settled, but Murray and Logan must have found it a wilderness. It is miraculous that they made such a good map, and it is quite evident that it was made by a strict use of the stratigraphic method that Logan had learned in England. Neglect of this stratigraphic method by a succeeding generation of geologists is probably responsible for the delay that followed in unraveling the geologic record during the next forty years.

Beneath this great series of sediments they found granite gneiss, which they recognized to be older and unconformable with the sediments. The name, Laurentian, had been given to similar granite gneiss along the Ottawa River, in what is here called the Grenville geological subprovince. In the Report of Progress for 1857, Murray applied this name, Laurentian, to the granite gneiss north of Lake Huron, thus implying the granite gneiss in the two regions to be of the same age. The sedi-

mentary series was at first called the Copper-bearing series, probably because copper had been found in it at Bruce Mines, but in 1858 Logan named it the Huronian. Huronian and Laurentian were together classified as Azoic, because no signs of fossils were found in them and also because they are overlain with marked angular unconformity along the coast of Lake Huron by sandstone and limestone that shelve gently southward and carry fossils like those in the European Ordovician.

SUBDIVISIONS INTO LOWER HURONIAN AND UPPER HURONIAN SERIES

The Huronian succession described by Logan and Murray became a standard for later investigators, especially those working in the South Shore sub-province south of Lake Superior. An area of especial interest to visiting geologists, in the neighborhood of Bruce Mines, where the most nearly complete sequence occurs, has become known as the Original Huronian area, because of the numerous visits and references to it. Geologists in the South Shore sub-province adopted the term Huronian for a similar succession south of Lake Superior. However, they discovered an unconformity within this succession and divided the Huronian into two series, Lower Huronian and Upper Huronian. This discovery led some of them to suspect that a similar unconformity must exist in Canada. Accordingly, numerous visits were paid to the Original area and the country eastward to Sudbury. Alexander Winchell, State Geologist of Minnesota, made several visits. In three of his resultant papers—those published in 1887, 1890, and 1891—he concluded from his own observations at various places between Echo Lake and Sudbury, and from the reports of other geologists, that the unconformity existed and lay between the lower and the upper slate conglomerates and below the limestone which lies between these conglomerates. Winchell's conclusion was afterward found to be incorrect, for the unconformity actually lies above and not below the limestone. Credit for discovery of the unconformity is due to Van Hise and Pumpelly. They found in a marble quarry, 2 miles northeast of Garden River railway station, the upper slate conglomerate in visible contact with the limestone, for a distance of a few yards. The conglomerate there contains many pieces of limestone up to a foot long, all identical in appearance with the limestone formation beneath. The line of contact follows rather closely the lamination of the limestone. These relationships were interpreted as evidence that the limestone had become a firmly cemented rock and had been subjected to erosion for a considerable length of time before the upper slate conglomerate was deposited upon it. They felt, however, that further deductions as to the nature and time value of the

unconformity could not be made at the time from one short outcrop. Curiosity on this matter had to wait upon the gradual exploration of other parts of the sub-province. It had not long to wait.

COBALT

In 1903 the spectacular Cobalt deposits were found, and for the next twelve years an area of about 5000 square miles was swarmed over by thousands of prospectors, geologically mapped, and mined in some places. The make-up of the country was simple. A series of sediments, 5000 feet or more thick and folded up to 45 degrees, lies upon a crystalline floor of granite, gneiss, and volcanic schists. Sills and dikes of quartz diabase, from which the silver-cobalt veins originated, intrude the older rocks. The unconformity between sediments and floor is the most conspicuous geologic feature in the region. The conglomerate formation at the bottom of these sediments is actually the upper slate conglomerate of the Original Huronian area, although this relationship was obscured by the absence underneath it of the lower series of the Huronian. The unconformity beneath the conglomerate series at Cobalt is, therefore, equivalent to the unconformity found by Van Hise and Pumpelly, plus the second unconformity at the bottom of the whole Huronian succession.

Although the Cobalt sediments were suspected to be Huronian, Miller decided upon a strictly local terminology for the Cobalt region. He called all the sediments the Cobalt series. These comprise two divisions quite unlike in appearance. Collins called the lower one, the Gowganda formation. This formation, about 4000 feet thick, is of scientific interest. It lies abruptly upon a fresh surface of schist, gneiss, and granite. Polished surfaces in this old floor are not uncommon and some doubtful glacial scratches have been recorded.[1] Boulder conglomerate like glacial till is the common phase of the formation and occurs mainly at the bottom. At least two scratched boulders have been found in it. Greywacke, the next most abundant rock, varies from massive to delicately banded or varved rock, and carries a few boulders and pebbles such as characterize glacial lakes, that could only have been dropped from floating pieces of ice. Some of the varved rock is locally crumpled. Quartzite is rare and impure. Geologists are well agreed that this assemblage was formed during an ancient Huronian glacial period.

[1] *See* A. P. Coleman [*A lower Huronian ice age*, Am. Jour. Sci., 4th ser., vol. 23 (1907) p. 187-192] *and*

M. E. Wilson [*Geology of an area adjoining the east side of Lake Timiskaming, Quebec*, Geol. Surv. Canada (1910) 46 pages].

The upper division of the series is a thick quartzite which Miller called Lorrain (sometimes Upper Cobalt to avoid double use of Lorrain or in the belief of an unconformity). It is a thick quartzite that descends, in some places abruptly, but usually gradually, into the Gowganda.

Miller called the quartz diabase sills and dikes, Nipissing diabase. The rock is uniform and widespread, being indistinguishable from the "greenstone" of Logan at Bruce Mines. It is remarkable for the tendency to develop basic and acid parts during crystallization and also for its valuable mineralizing effect in producing deposits of silver-cobalt, copper, and gold.

ANIMIKIE SEDIMENTS AT SUDBURY

The great nickel-copper sulphide deposits of Sudbury have had a much slower development, economically and scientifically, than the silver-cobalt field. Except for a small ore production at the subsidiary camp, Gowganda, Cobalt has been "dead" since 1935. Sudbury, on the other hand, produces more ore each year. The ore was discovered in 1856 by Land Surveyor A. P. Salter, and its character was determined by Sterry Hunt, of the Geological Survey, from specimens collected by Alexander Murray. But prospectors did not arouse to the importance of the discovery until 1884, when ore was exposed in a rock-cut of the Canadian Pacific Railway, then under construction. Development was also delayed because the metal, nickel, had been discovered comparatively recently, commercial uses for it had not been found, and its presence with copper presented entirely new problems to the metallurgists.

Mining operations in the Sudbury field have made it a center for intensive geological study. Besides the great interest of its ore deposits, the field has yielded several important additions to the geological history and succession of the Timiskaming sub-province. One of these discoveries was a new series of volcanic sediments. Bell (1891) found an elliptical, basin-shaped area, 37 miles long, underlain, in ascending order by a thick volcanic breccia, dark gray to black, fine-grained slate, and a dark gray, feldspathic sandstone. This series is so much less folded and recrystallized than the Huronian rocks in neighboring parts of the area that Bell was inclined to regard them as unconformably above the Huronian and either Upper Huronian or Lower Cambrian in age. The distribution and succession of the various formations are represented in Map No. 130, Sudbury Sheet (now out of print) on a scale of 1 inch to 4 miles, issued by the Geological Survey of Canada.

Coleman began his famous survey of the Sudbury mining field for the Ontario Bureau of Mines in 1902. His report and map contain a

highly detailed description of Bell's so-called Cambrian or Upper Huronian series. He found the series just to fill the spoon-shaped concavity of a great laccolithic sheet of norite-micro-pegmatite, the ore-bringing formation. It consists of four conformable divisions. At the bottom, according to Coleman, is the Trout Lake conglomerate, a coarse agglomerate, with some rounded boulders, about 500 feet thick. This grades imperceptibly upward into the Onaping tuffs, 3800 feet thick, of tuffs and lesser proportions of allied volcanics. The Onaping tuffs grow finer from bottom to top. The uppermost part is a black carbonaceous material that merges imperceptibly into 3700 feet of finely stratified carbonaceous Onwatin slate. The Onwatin slate gives place, in turn, by a few alternating beds, to the Chelmsford sandstone, a feldspathic sandstone or poorly cemented quartzite of angular texture that contains enough carbonaceous matter to color it a sooty gray. Erosion has destroyed the upper part of the Chelmsford sandstone, leaving a greatest thickness of 1500 feet. Besides the disseminated content of carbon, the series is intersected by veins from paper-thinness to 6 to 8 feet thick of a jet-black hydrocarbon, anthraxolite, which contains as much as 96 per cent fixed carbon. It is supposed that the Onwatin slate was originally an oil shale and that the anthraxolite veins are fissure accumulations of oil from which the volatile part evaporated. The Animikie or Upper Huronian on the northwest side of Lake Superior contains carbonaceous slate and anthraxolite veins exactly similar to those at Sudbury; in fact, it is the only other known occurrence of the kind in the Lake Superior-Lake Huron region. Both series are much less folded and indurated than any of the neighboring pre-Cambrian rocks, and they both lie in the upper part of the pre-Cambrian succession. On the strength of these resemblances, Coleman classified the series at Sudbury as Animikie. The series is entirely unlike any part of the Original Huronian. The determination of its position in the pre-Cambrian geological column forms a problem that is still unsolved.

This paper deals mainly with the geological history of the Timiskaming sub-province and only incidentally with the mineral deposits. But Sudbury is such an important mining field that a few words more than usual about it may not be out of place. A great lacolithic body of igneous rock in the form of the bowl of a spoon is the source of the ore. The ore bodies are distributed along the outer rim or bottom and along dike-like extensions, called offsets. The ore consists almost entirely of pyrrhotite, chalcopyrite, and pentlandite. Its chief metallic contents are nickel and copper, but it also contains recoverable amounts of platinum, palladium, osmium, rhodium, ruthenium, selenium, gold, and silver,

besides great quantities of sulphur and iron. The magnitude of production may be judged by comparison of the values of the products recovered in 1890, the first year of recorded production, and in 1935:

Product	Production		
	1890 (pounds)	1935 (pounds)	1935 (ounces)
Nickel	356,000	140,604,207	.....
Copper	3,322,000	238,139,392	.....
Platinum	.....	.....	128,874
Gold	.....	.....	69,944
Silver	.....	.....	3,160,222
Selenium	.....	72,616	.....
Tellurium	.....	9,987	.....

RE-STUDY OF THE ORIGINAL HURONIAN AREA

GENERAL STATEMENT

When the Twelfth International Geological Congress met in Canada in 1913, great interest was manifested by visiting geologists in the stratigraphy and geological history of the pre-Cambrian, especially certain critical localities, of which the Original Huronian area was the chief. There was, however, little information available about this key area beyond what had been obtained by Logan and Murray, 65 years before. The deficiency was rendered so obvious that, the next year, the Geological Survey of Canada undertook a review, lasting from 1914 until 1917, of the country north of Lake Huron, from Sault Ste. Marie to Sudbury, in which Logan and Murray had first explored the Huronian formations. Three key areas were mapped in considerable detail, and enough work was done in the intervening spaces to assure reliable correlation. At Sudbury this belt joins with the Huronian area extending into Quebec which had been explored as a result of the discovery of the silver-cobalt ore deposits in 1903, so that continuous correlation across northeastern Ontario was at last accomplished, and the compilation of a geological map of the whole region was rendered possible for the first time.

Some changes and additions were made to the Huronian succession as defined by Logan and Murray. The revised succession, inclusive of

all changes up to 1935, and the original one by Logan and Murray are here placed side by side for easy comparison:

Collins (1914–1917)	Feet	Logan and Murray (1845–1858)	Feet
Cobalt Series:			
White quartzite	400+	White quartzite	400
Yellow chert and limestone	200	Yellow chert and limestone	200
White quartzite	1500	White quartzite	1500
Banded chert quartzite	700	Yellow chert and limestone	400
		White quartzite	2970
Lorrain quartzite	5500–6500	Red jasper conglomerate	3000
		Red quartzite	2300
Gowganda formation	0–3500	Upper slate conglomerate	3000
Angular and erosional unconformity		No unconformity recorded	
Bruce Series:			
Serpent quartzite	1100+	(Not recorded)	
Espanola limestone and limy silt	400–700	Limestone	300
Bruce conglomerate	20–500	Lower slate conglomerate	1280
		White quartzite	1000
Mississagi quartzite	1000–8000	Chlorite slate	2000
		Grey quartzite	500
Ramsay Lake Conglomerate	0–250	(Not recorded)	
Major angular and erosional unconformity		Unconformity	
Pre-Huronian granite-gneiss, sediments and volcanics		Laurentian gneiss	

RAMSAY LAKE CONGLOMERATE

Logan and Murray do not mention a conglomerate at the bottom of the Huronian. It was discovered by Irving (1887) on an islet in Lake Huron, 3 miles east of Thessalon. This place was visited later by Pumpelly and Van Hise (1892), who give an excellent description of the conglomerate as seen on the islet and on the nearby mainland. The pre-Huronian basement at these places is granite gneiss. Beginning 10 or 15 feet away from the gray quartzite that overlies most of it, this gneiss shows the first signs of weathering. It is fractured and composed of angular pieces that have not moved out of place and are separated by narrow crevices filled with coarse grit. In the succeeding 5 feet, the blocks moved a little out of their original positions, became less angular and with larger interstitial, grit-filled spaces, and finally merged into a boulder conglomerate containing a variety of rock types. This conglomerate is only 10 or 15 feet thick and grades upward into feldspathic quartzite. It is evidently a thin bouldery soil formed by disintegration and decay of the pre-Huronian floor and slightly worked to form conglomerate.

Collins and his associates found this conglomerate at intervals from Thessalon to Sudbury, where it had been observed by Coleman and named Ramsay Lake conglomerate because it is well exposed on the shore of a lake of that name just outside Sudbury. The conglomerate varies greatly from place to place but was identified by its position between the Mississagi quartzite and the pre-Huronian and by its characteristics of an ancient pre-Huronian soil. It varies in color and in composition with the local nature of the underlying pre-Huronian. Where the latter is granite gneiss, the matrix of the conglomerate is light-colored arkose, and boulders of granite gneiss are likely to predominate. Where volcanic schists form the pre-Huronian floor the matrix of the conglomerate is a dark greenish chloritic mixture, and pebbles of the volcanics are more abundant than other rock types. Likewise, over pre-Huronian sediments (Sudbury series) the Ramsay Lake conglomerate partakes of the characteristics of these sediments. This resemblance to the locally underlying pre-Huronian rocks is most pronounced where the conglomerate is thin and where it grades downward into the undisturbed decayed top of the pre-Huronian basement. Thick parts of the conglomerate are likely to be more uniform and to consist of a great variety of pebbles and boulders, up to 5 feet across, in an abundant matrix, rather light gray or brown, composed of coarse particles of quartz, decayed feldspar, and various unidentified secondary minerals.

The conglomerate ranges in thickness from nothing to about 250 feet, but, according to R. C. Emmons, it becomes far thicker and more diversified stratigraphically on the north side of the Huronian area, from Garden River eastward for 20 miles or more. Emmons is the only geologist who has described this unusually thick portion of the conglomerate. A review of his work should be made, especially as R. G. McConnell found sediments farther west, near Garden River, which he considers pre-Huronian and worthy of the name Soo series.

Although the lower contact is usually indefinite within 6 inches to 3 feet, in consequence of pre-Huronian weathering, there are places where the junction with the pre-Huronian is a sharp line and no weathered products remain in place. The two kinds of contact may occur close together, as at the original locality, 3 miles east of Thessalon.

At Ramsay Lake, Coleman and others claim that the conglomerate contains boulders and pebbles of the Mississagi quartzite and is, therefore, unconformably above the quartzite. Collins, from observation of many contacts, believes that the phenomena are due to crushing and faulting and that the Ramsay Lake conglomerate forms the base of the Mississagi quartzite. Numerous examples of similar crush breccia

have been seen in the Mississagi quartzite between Sudbury and Wanapitei Lake.

MISSISSAGI QUARTZITE

Alexander Winchell (1888) gave the name Mississagui to the Huronian quartzite, in the Original Huronian area, that lies below the lower slate conglomerate. In 1914 the spelling was changed to Mississagi, to conform with the ruling of the Geographic Board of Canada. At that time, the Ramsay Lake conglomerate, which had been seen only at a few places where it is thin, was included in the Mississagi.

The Mississagi is mainly a uniform feldspathic quartzite, which thickens southward from 1000 feet to about 8000 feet in the country south of Sudbury. Visiting geologists will find it well exposed along the coast west of Blind River and on Panache Lake. Collins at first estimated the extreme thickness at 12,000 feet, owing to its steep and apparently monoclinal attitude across great widths, but further mapping has indicated a repetition of beds by close folding. The quartzite is in beds up to 5 or 6 feet thick, in many places separated in the lower part of the formation by thin argillitic partings that weather into grooves. This variety is strikingly displayed near Wanapitei on the trans-Canada highway. Within each bed it is strongly and abundantly cross-bedded. Symmetrical ripple marks, some as much as 6 inches in amplitude, are also common. The quartzite is nearly white, coarse grained, and firmly cemented. Where the outlines of original sand grains can be distinguished, they are not well rounded. A fine silty material composed of secondary mica from decayed feldspar fills the interstices. There seems to be some upward diminution in the percentages of this interstitial material and also of feldspar, especially in the first few yards from the bottom.

Thin lenses of conglomerate occur locally and sparingly at different horizons. They are composed of well-worn pebbles of granite or quartz, not more than 2 inches across, in a quartzite formation like the Mississagi. However, one conglomerate was found on the north side of Lake of the Mountains, near Blind River, which is not so easily explained.[2] It is a coarse boulder conglomerate, crowded with well-rounded boulders of numerous kinds, up to 2 feet across. Only a powerful stream or a freshet could have carried such large masses of rock.

The Mississagi formation also contains a few bodies of delicately stratified, thoroughly weathered argillite. These are probably local accumulations of limited horizontal extent; they occur at various horizons from the base upward. They attain thicknesses of several hundred feet.

[2] W. H. Collins: *The north shore of Lake Huron*, Geol. Surv. Canada, Mem. 143 (1925) p. 43.

Deposition of the Mississagi in the shoreward parts of a large lake or sea, in water up to 30 feet or more deep, is indicated by the thick beds, cross-bedding, broad ripple marks, and beach-like conglomerate beds. Apparently, the basin of deposition deepened in a southerly and easterly direction, for the formation thickens greatly in these directions.

The chloritic slate of Logan seems to refer to an extensive lava flow near Thessalon of a dark-green rock containing about 50 per cent silica.[3] This flow is amygdaloidal and also ellipsoidal in places. It lies directly against the lower part of the Mississagi quartzite and seems in places to bevel the bedding planes of the quartzite. Whether it is contemporaneous with the quartzite or lies on an erosion surface has not been ascertained. A similar basic flow, now much sheared and recrystallized by younger (Killarnean) granite occurs along the Canadian Pacific railway between McFerson and Cutler stations. This occurrence would place the time of eruption between Mississagi and Killarnean. A little ellipsoidal greenstone is also associated with Mississagi quartzite south of Birch Lake. Some ellipsoidal green lava also occurs with the quartzite south of Birch Lake. Evidently, a slight vulcanism occurred in or about Mississagi time.

BRUCE CONGLOMERATE

The Mississagi quartzite passes abruptly upward into a boulder conglomerate of utterly different appearance. Passage from one to the other varies from a gradation of 40 to 100 feet at one extreme to a perfectly sharp line, suggesting disconformity. The lower slate conglomerate of Logan, or Bruce conglomerate, as re-named in 1914, is an unstratified boulder conglomerate, consisting typically of subangular to round pebbles and boulders, up to 3 feet across, in a dark-gray matrix. Boulders form one-third to one-half the volume of the rock in the lower part and grow scarcer toward the top of the formation, where the almost pebble-free matrix grades into a thinly stratified siliceous silt that forms the base of the Espanola formation. Locally, in its upper part the conglomerate shows stratification in the form of a few pebble-free beds, a few feet thick.

There is a striking resemblance between the Bruce conglomerate and the unstratified boulder conglomerate of the Gowganda formation, which naturally raises the question: Is the Bruce conglomerate also a glacial tillite? Both resemble glacial till in the great many kinds of boulders, the presence together of soft and resistant rock types, and the high proportion of matrix to inclusions. In other respects, however, they differ a good deal. Striated boulders have not been found in the Bruce

[3] *Op. cit.*, p. 76.

conglomerate, though this may be due to the greater difficulty of freeing boulders from the matrix, as compared with the Gowganda conglomerate. Unlike the Gowganda the Bruce does not rest unconformably and in knife-sharp contact upon a smooth pre-Huronian pavement, but grades conformably downward into the Mississagi quartzite. There is no varved greywacke associated with the Bruce conglomerate.

Also, the cementing materials differ considerably. That of the Gowganda tillite is a homogeneous slate-gray rock flour composed of fresh mineral particles. The matrix of the Bruce conglomerate is coarser and less homogeneous. It consists of angular to subangular grains of quartz, and fresh feldspars, up to 2 millimeters in diameter, in a finer paste of chlorite, sericite, epidote, small particles of quartz and feldspar, and some opaque indeterminate matter. The coarse grains form one-fourth to one-half the matrix. Those of feldspar are fresh and comprise both potash and soda varieties. Scattered grains of pyrite are a rather characteristic feature. Another is a tendency of the matrix to shrink away from the pebbles and boulders, leaving narrow crevices around the latter.

The Bruce conglomerate is much too poorly assorted to have been deposited in water, nor could it have been transported far by water. Much more likely, it was deposited on land in the form of a fluid, probably mud flow, too thick to permit of assortment of its constituents.

There is a definite and fairly regular thickening of the formation southeastward. It is only 20 feet thick at Echo Lake, but on Lake Panache it reaches somewhere between 500 and 800 feet.

ESPANOLA FORMATION

Logan and Murray evidently saw only the lower part of this formation, which contains the largest proportion of carbonate. In 1914, it was divided into three parts: a lowermost one, composed mainly of carbonate, about 300 feet thick, which N. H. Winchell (1891) named the Bruce limestone; then a siliceous one, nearly devoid of carbonate and 300 to 400 feet thick, named the Espanola formation, which is well exposed near Espanola; and at the top, another carbonate-rich, red-weathering portion, 0 to 75 feet thick, named the Espanola limestone. During later work in the general neighborhood of Sudbury, it was found impracticable to distinguish these three divisions, which are really carbonate-rich and carbonate-poor variations of one continuous formation, so the name Espanola formation has been adopted for all three.

Essentially, the Espanola formation is a mixture of silt and magnesian limestone more or less well separated into alternating layers, from a fraction of an inch to a foot thick. Weathered surfaces are strongly corrugated, owing to the unequal resistance of the siliceous and the cal-

careous layers. Siliceous parts of the formation also weather into innumerable lenticular crevices, an inch or two long and diagonal to the bedding, which suggest torsion cracks; they are highly characteristic of the Espanola formation. Ripple marks, current marks, and mud cracks occur in the siliceous layers. The carbonate beds are pale gray and red-weathering near the top of the formation; the siliceous layers vary from light gray to fairly dark gray-green.

The siliceous material is silt, consisting of sharply angular grains of quartz and feldspar in a paste or matrix of chlorite, sericite, and biotite shreds, minute particles of quartz and feldspar, and some opaque white matter. Potash and soda feldspar are present, both quite fresh. The large grains of quartz and feldspar range in size from .1 to .75 millimeter, and their volume proportion to the paste varies almost as greatly. Where not affected by granitic intrusions the carbonate portion is a fine-grained aggregate. Some of it is disseminated through the silt, but the great bulk is in fairly pure layers, with clean-cut edges. Chemically, it is a mixture of the carbonates of lime, magnesia, iron, and possibly a little manganese. Iron carbonate is negligible except near the top of the Espanola formation. Magnesium carbonate varies from two-thirds of the lime carbonate content to less than one per cent of it.[4]

Owing to the mixture of carbonate and hydrous secondary silicates the Espanola formation is easily metamorphosed. In the neighborhood of Panache Lake, where intrusions of younger granite are abundant, it is more or less recrystallized to biotite, scapolite, tremolite, and other lime silicates. The extreme product is a black hornfels in Killarney Bay.

The almost pebble-free top of the Bruce conglomerate grades, in from 2 to 10 feet, into a thin laminated silt that forms the base of the Espanola. At the top of the Espanola, there is a more gradual change from the laminated silt into a delicately stratified, fine-grained feldspathic quartzite that represents the base of the Serpent quartzite. Within the Espanola, a few exposures of thin beds of conglomerate are believed to represent local emergences and interruptions of deposition.

At one place, in timber berths 149 and 155, there is an iron formation, 2½ miles long and about 15 feet thick, just above the base of the Espanola. It consists of laminae of weakly magnetic black iron ore alternating with laminae of silt. Probably, it represents chemical or organic precipitation of iron solutions that collected in a sheltered lagoon or bay of the larger body of shallow water in which the Espanola formation was deposited. The Espanola formation is quite plainly a product of shallow-water deposition. In places and at times, it became dry land.

[4] *Op. cit.*, p. 50.

One must think of it as occupying a flat, featureless, marshy region, fed by streams or winds carrying fine, silty dust.

SERPENT QUARTZITE

The Serpent quartzite was first recognized in 1914 and was so named because it is well exposed at the headwaters of Serpent River. There is a gradual change from Espanola silt to Serpent quartzite, with recurrences of the silt until, in 15 to 200 feet, a uniform fine-grained white quartzite is reached. This quartzite differs from all others in the Huronian in having a texture suggestive of unglazed porcelain. Also, it possesses an extremely delicate paper-thin lamination that can be seen only on weathered surfaces, where it is expressed as a minute corrugation. Microscopic examination does not reveal the cause of this structure. Where the outlines of the original sand grains are discernible, they are subangular to round. The feldspar grains are fresh. A little disseminated carbonate is common.

There are about 1000 feet of this fine-grained variety, and, in most places, it ends in an erosion surface. South of Panache Lake, however, it grades up into a coarser pink feldspathic variety, and reaches a total thickness of possibly 4000 feet. A few thin beds of quartz pebble conglomerate were seen well up in the formation.

BRUCE-COBALT INTERVAL

Pumpelly and Van Hise based the existence of an erosional unconformity within the Huronian on a single outcrop near Garden River, where the upper slate conglomerate of Logan rests on the limestone. Much more has been learned about this unconformity since then. Throughout the area of 5000 square miles or more around Cobalt that was prospected for silver-cobalt ore deposits and geologically mapped between 1904 and 1920 the upper slate conglomerate (Gowganda formation) was found to lie in conspicuous angular unconformity upon the pre-Huronian basement. The Bruce series (Lower Huronian) does not intervene in this part of the Timiskaming sub-province. During 1914 and 1915 the unconformity was examined in the Huronian area between Sault Ste. Marie and Sudbury, where the Bruce series does underlie the Cobalt series. Plenty of evidence of an erosional unconformity was found. Just west of Echo Lake the conglomerate rests upon 40 feet of Espanola limestone, and, a quarter of a mile away, it rests on Mississagi quartzite, indicating a beveled erosion surface. There is an inch or two of weathered material at the top of the limestone; otherwise, the contact is quite sharp. For the first one to three feet the conglomerate contains few inclusions, but above that it carries angular pieces of the limestone

and some well-worn pebbles of quartzite. On the shore of Lake Huron, near the west side of Plummer Additional Township, the conglomerate lies in knife-sharp contact on an undulatory surface of Serpent quartzite.[5] Worn pebbles of the characteristically laminated Serpent quartzite occur in the conglomerate above another similar contact, half a mile inland. Practically the same relations were seen on an islet in Long Lake, Timber Berth 150. A wavy clean-cut contact of the Gowganda conglomerate with Mississagi quartzite on the west side of Chiblow Lake may also be mentioned to indicate at what widely different horizons in the Bruce series the unconformity occurs.

At all the mentioned localities the contacts are sharply defined. In many other places, however, the contact might readily be mistaken for a conformable one, without interruption of sedimentation. In parts of Timber Berths 137, 138, 143, and 144 the Serpent quartzite passes gradationally, in about 15 feet, into Gowganda conglomerate or, in the same thickness, it alternates with thin beds of pebble conglomerate, which assume successively a more bouldery appearance and finally merge into the great boulder conglomerate of the Gowganda.

"The relationship is so positively one of unconformity in some places, and so like conformity in others, that it was finally decided to trace the contact between the two series continuously and with great care for a considerable distance, to see if it does not bevel two or more of the Bruce series formations. This was carried out . . . beginning near the middle of township 155 and continuing across township 149. At the west end of this section of contact the Gowganda conglomerate overlies the Bruce limestone, never less than 20 feet above the base of the limestone and never more than 75 or 100. Where actually exposed the contact . . . is sharply defined, and angular fragments of limestone are abundant in the first 10 feet of conglomerate. . . . This relation persists eastward for several miles. Finally, south of Burnt Lake the contact rapidly ascends across the Bruce limestone and Espanola formation to the Serpent quartzite, in which formation it remains all the way around to McLeod Lake."[6]

As the Cobalt series rests on the Serpent quartzite at one extreme and the Mississagi quartzite at the other extreme, erosion must have cut into the Bruce series as much as 1700 feet deeper at some places than at others. This is the differential amount of erosion. How much greater the whole amount was is not determinable. Judging from the way the contact follows one horizon for a long distance and then changes abruptly to another, the pre-Cobalt surface of the Bruce series must have been mesa-like, with alternating escarpments and level areas, a type of landscape not associated with glaciation.

[5] W. H. Collins: *The north shore of Lake Huron*, Geol. Surv. Canada, Mem. 143 (1925) pl. 13.
[6] *Op. cit.*, p. 72.

It is rarely possible to measure the angle of unconformity between Bruce and Cobalt series, because the base of the Cobalt series is usually an unstratified or poorly stratified boulder conglomerate. From Sault Ste. Marie eastward to Lake Panache, no discordance is perceptible on the maps. That a small angle does exist, however, was seen in the southwest bay of Quirke Lake, where an angle of about five degrees occurs between the bedding planes of the Serpent quartzite and some sandy beds in the Gowganda boulder conglomerate just above. Between Panache Lake and Wanapitei Lake the Gowganda formation is too scantily represented for the angularity of the unconformity beneath it to be observed. East of Wanapitei Lake, however, the angle between the Gowganda formation and the Mississagi quartzite, Bruce conglomerate and Espanola formation on which it lies is large—probably as great as 45 or 50 degrees. Whether this is a local condition or represents a gradual and regional increase of folding of the Bruce series in pre-Cobalt times is conjectural. If it were regional, one would expect to find remnants of downfolded Bruce series for a considerable distance north and east of Wanapitei Lake, but these have not been recognized. The Bruce series appears either never to have extended far in these directions or to have been eroded before the Cobalt series was deposited. If a good survey were made of the eastern half of the Wanapitei one-mile sheet of the Canadian Geological series, much light would be cast upon the structural form of the Cobalt-Bruce surface.

The Bruce series appears to have experienced some deformation while it was being laid down. For instance, the Espanola silt is crushed in many places into a coarse breccia, composed entirely of its own hardened fragments. Again, the whole Bruce series, from the conglomerate upward, is cut by peculiar sandstone dikes that are attributed to earthquakes. These forces were probably small and did not much affect the shape of the Cobalt-Bruce surface. All in all, the surface was probably only mildly deformed.

GOWGANDA FORMATION

By all means the most interesting formation in the Huronian is the Gowganda, or upper slate, conglomerate. It is a slate-gray formation of glacial origin, from 3000 feet to 4000 feet thick. Its commonest rock is a coarse boulder conglomerate, or tillite, usually found at the bottom of the formation. It lies upon a remarkably flat and fresh surface of pre-Huronian. At least two striated boulders have been found. Such other unusual materials as perfectly fresh pyrite pebbles occur in it in a few places. An unstratified thick-bedded greywacke is common. In many places, this greywacke or rock flour becomes seasonally banded (varved) in the most delicate and continuous manner. Small boulders

and pebbles in it appear to show that they were deposited in the clay from pieces of floating, melting ice. Local crushed material also indicates freezing and thawing action.

LORRAIN QUARTZITE

The Gowganda formation is succeeded by a great thickness of quartzite. One formation grades more or less quickly into the other. The lower part of the Lorrain is highly feldspathic and ranges from pale sea-green to dull red. About 3000 feet up in the formation, small lenses of quartz conglomerate appear. Near Bruce Mines in the Original Huronian area, some of these pebbles are bright red, giving the rock a brilliant appearance. This is the red jasper conglomerate of Logan. Elsewhere in northeastern Ontario the quartz pebbles are white or gray. Above the conglomerate is another pure white quartzite, 2000 or 3000 feet thick. This is quarried at Killarney, where it occurs next the coast and is shipped to Niagara Falls and other industrial plants for making ferro-silicon and other products.

CHERTY QUARTZITE

About 700 feet of fine-grained and finely stratified, brown and yellow quartzite overlies the Lorrain. It contains some carbonate and is a typical silt, composed of sharp particles of quartz and feldspar mingled with secondary mica.

STRATA ABOVE CHERTY QUARTZITE

The strata above the Cherty Quartzite have been seen only by Logan and Murray, but their descriptions are probably reliable.

KILLARNEAN GRANITES

In the foregoing re-examination of the Huronian, it was found that these formations became more closely folded eastward from the Original Huronian area. Dips of 45 degrees, or less, increase to verticality in the neighborhood of Lake Panache. In this direction, also, dikes and other small bodies of intrusive granite begin to appear. At the east end of Lake Panache, these granitic bodies united to form a mass of batholithic dimensions. It was apparent that the Huronian had been mountain-built.

These granites intrude the Nipissing diabase, producing in the latter a contact zone full of garnets. They also intrude the other Cobalt and Bruce formations as far down stratigraphically as the Mississagi quartzite.

At first, the granites were found around Lake Panache, at Killarney, and westward along the coast of Lake Huron as far as Serpent, Indian

Reserve. They are mostly deep salmon color, with large crystals of feldspar and little quartz. Another variety, seen on Lake Panache and at Serpent, is finer-grained and gray. Chemical analyses show an extremely close resemblance, the most characteristic common feature being that the potash content is generally greater than the soda, in the ratio of about three to two, and the total alkali content is high. Logan was the first to observe this younger granite. In the *Geology of Canada* (1863), he refers to a plateau-like area south of Lake Pawkowkanii, where he found the older gneiss cut by many stringers and irregular bodies of granite. The *Atlas* also shows a dike of syenite near Batchawana Bay that cuts the Huronian formation. Barlow affirmed that the granite at Killarney cut some of the Huronian formations there.

Bell,[7] with whom Barlow was associated, gave the Indian name, Thebahnahawaning (meaning Little Current) to Killarney, but the latter name eventually was given to the granite. This name, Killarney, was adopted for all the younger granites along the north shore.

The age of the granites along the coast could be ascertained fairly well, because they intersect the whole succession of Huronian sediments. Later, granites such as the Birch Lake, the Creighton, and the Murray at Sudbury, and others as far west as Sault Ste. Marie, were placed with the Killarney granites, although their age relationships were not so closely determinable. This gave rise to the looser term Killarnean. It has been found impossible, for example, to ascertain whether the Creighton and Birch Lake granites are older or younger than the nickel irruptive and the Whitewater series.

If the closely folded Huronian sediments and the intrusive Killarnean granites represent a period of mountain-building, and if this mountain-building was followed by peneplanation or other prolonged erosion, the surface of the erosion would constitute a suitable top for the Huronian system or whatever other great time division it should be called, just as the great pre-Huronian unconformity forms the bottom. However, the amount of erosion after the Killarnean is not yet measurable. Some uncertain light has been thrown on this matter by investigations that are described briefly in the next section.

SUDBURY

Small occurrences of lead, zinc, pyrite, and other materials had been known for thirty years to exist here and there within the Sudbury nickel basin. These had been considered of little value, but, in 1925, a pros-

[7] Bell believed that the Indian language should be preserved as much as possible by giving Indian names to lakes and other geographic features. An extreme example of this practice is a small lake in the northeast corner of Geological Survey of Canada Map 130, which rejoices in the name of Ashiginipoonsapoganing, though on more recent maps it has been shortened to Ess Lake.

pecting syndicate undertook to diamond-drill one of them, situated near the great bend of the Vermilion River. It found that the mineralization occurred along or close to a series of faults that extended along the major axis of the basin and passed out of the basin at its southwest end as two large faults.

The chief mine development was the Errington. This led the Ontario Department of Mines to re-study with great care the Animikie series of Coleman, or the Whitewater series as it had been afterward named. The Whitewater series snugly fills the inside of the Nickel basin, which is generally regarded to be of the form of an oval saucer. It does not extend outside the basin. It is concentrically stratified almost exactly after the shape of the inner margin of the irruptive. The color ranges from dark green to gray to black. A secondary schistosity, which varies from 75 degrees on one side of the folds to 75 degrees on their other sides, affects the whole series, but especially the sandstone and the slate. This schistosity strikes along the major axis of the basin or at small angles to it.

The bottom of the series, for as much as 500 feet in thickness, is an extremely coarse aggregate of blocks and boulders of various igneous rocks. Coleman also reports a small percentage of granite pieces, and, on this account, he called this part of the series the Trout Lake conglomerate. Thomson reports having seen raft-like pieces of quartzite as much as 100 feet long. Burrows and Rickaby concluded that the separate name for this division was not needed, and they included it in the main body of the overlying Onaping tuff. The formation is, at the base, a decidedly coarse and angular tuff or agglomerate, composed of pieces a foot or more in diameter and including a great variety of volcanic materials, and even some sediments. Proceeding upward, the formation grades finer rapidly, and the main part of the formation is a nearly black tuff or ashrock.

Near its junction with the overlying Onwatin slate the Onaping formation becomes quite black, owing to a small content of carbon. The change from one formation to the other is a matter of opinion, but the closest judges would probably not locate it within 10 feet. The Onwatin slate is a uniform formation throughout the Nickel basin. It is almost black, finely laminated, and has slaty cleavage. It contains from three to about fifteen per cent of carbon, evidently most abundant in the lower part of the formation.

In a few places, particularly in Baldwin Township, the slate contains small veins of a black shiny material like jet, that has been called anthraxolite. The mineral contains as high as 96 per cent of carbon, the remainder consisting of sulphides and quartz. These veins have been

considered at several times as a source of anthracite fuel, but they are much too small to be mined for this purpose. According to geological opinion, the Onwatin slate was once an oil shale, and the present anthraxolite veins represent cracks in it, which were filled with oily matter that gradually dried out to a solid material like jet.

The slate passes upward by alternating beds, in 10 or 15 feet, into the Chelmsford sandstone. This rock is feldspathic quartzite, nearly black, owing to the presence of a small percentage of carbon. It was deposited in beds, 4 to 8 feet thick, and has since been greatly sheared. The rock contains almond-shaped iron-bearing concretions, which weather out rather prominently, but, otherwise, the formation appears to be of little economic or scientific interest. Erosion has left not more than 1500 feet of it. Estimates of the thickness of the series differ as follows:

	Trout L. (feet)	Onaping (feet)	Onwatin (feet)	Chelmsford (feet)	Total (feet)
Coleman.....	500	3800	3700	0–800	.....
Collins and Kindle...	0	4500–5000	2500	0–1500	.....

Why is there a solitary area of Whitewater series in the whole region of northeastern Ontario? The original deposit, being volcanic, may never have been extensive. If the series had been deposited widely before the Killarnean mountain-building, it would have been infolded at various places and preserved from erosion as strips. But this is not the case. If the series had been widely deposited upon a peneplaned or other surface of mild relief, it would be likely to be eroded away. Several patches of nearly horizontal sediments and volcanics on the east coast of Lake Superior, which are generally regarded as Keweenawan, lend a little support to this third theory. Recent work on the Sudbury nickel basin also lends some support to the idea of a peneplain beneath the Whitewater series. The transition zone, which lies between the norite and the micropegmatite layers of the irruptive, was found to be strictly uniform in density at 2.70 to 2.82. The transition zone is also closely concentric with the norite and the micropegmatite. To explain these two features, it appears necessary to assume that the transition zone was in a strictly horizontal plane when it crystallized and that, therefore, the whole nickel irruptive was also a horizontal body, although it is now shaped like the bowl of a spoon. If this were so, the Whitewater

series of sediments must also have lain horizontally when the nickel irruptive was intruded beneath them. It is worth remembering, also, that the Animikie in western Ontario rests upon a profound unconformity.

If such an erosional surface exists in northeastern Ontario below the Whitewater series, all three features—mountain-building, batholithic invasion, and peneplanation—mark the top of the Huronian.

MISCELLANEOUS

TRAP DIKES

In the region of Sudbury, there are numerous dikes to which the name trap has been given. These intersect all formations up to, and including, the Nickel irruptive. They strike fairly uniformly in a direction slightly north of east. Mineralogically, they consist mainly of hornblende and albite. The chemical composition is indicated by the following two analyses:

Constituent	Per cent	
SiO_2	51.12	52.54
TiO_2	1.05	
Al_2O_3	16.03	16.10
Fe_2O_3	1.58	5.88
FeO	11.91	9.94
MgO	4.56	1.66
CaO	8.75	7.35
Na_2O	2.55	3.02
K_2O	1.24	1.58
H_2O+	0.51	1.84
H_2O-	0.10	.....
MnO	0.15	.....
P_2O_5	0.19	.....
BaO	0.06	.....
NiO	None	.....
Cr_2O_3	None	.....
Cl	Trace	.....
CO_2	None	Trace
S	0.36	.....
Total	100.14	.....
Less O/S	0.27	.....
	99.87	99.91

The first of these samples was collected by T. C. Phemister, from a 40-foot dike, cutting the Birch Lake granite in lot 7, con. 3, Shakespeare Township, and was analyzed by R. J. C. Fabry, of the Geological Survey of Canada. The second sample is from the Creighton mine. According to Phemister, the rock is luciite.

OLIVINE DIABASE

The whole region of northeastern Ontario, from Lake Superior to Lake Timiskaming, is intersected by many thousands of olivine diabase dikes. These dikes are probably the youngest formation in the pre-Cambrian; they cut the trap dikes already mentioned. The rock is remarkably fresh, its constituents, labradorite, augite, olivine, and black iron ore, being almost unaltered. It ranges in chemical composition from about 45 per cent to 50 per cent silica. Characteristic features of many of these dikes are phenocrysts of feldspar up to two inches in diameter, which occupy the central portion of the dikes. These dikes are not only uniform in composition but they tend to strike in nearly parallel directions, the direction varying throughout the northwesterly and southeasterly quadrants. In some places, although the rock is quite fresh, it disintegrates into blocks and granular sand, the whole resembling crumbled masonry. No parent mass of these dikes has ever been discovered, except one small mass in Flett Township, Ontario.

KEWEENAWAN FLOWS AND SEDIMENTS

On the east side of Lake Superior, there are two small areas of sandstone and amygdaloidal volcanics, which are commonly accepted as extensions of the great Keewanawan series that occupies Keweenaw peninsula. These small areas are mere erosion scales lying on the older pre-Cambrian rocks at angles of 20 degrees or less. They are significant in that they appear to lie upon a much more ancient floor, possibly a peneplain. This adds some strength to the belief that an unconformity of major order lies between the Killarnean granites and the Whitewater series.

PALEOZOIC COVER

The relationships of the Paleozoic formations to the pre-Cambrian Shield are best displayed along the north coast of Lake Huron, from Killarney to Sault Ste. Marie. The Paleozoic beds shelve off the pre-Cambrian at gentle angles, indicating an underlying surface with gentle to hilly relief. In the eastern and central parts, the Paleozoic rocks are of Ordovician age, chiefly Beekmantown, and are prevailingly gray. Close to Sault Ste. Marie, they change to the St. Mary's sandstone, probably Cambrian. The main part of this formation is a deep brick-red sandstone, sprinkled with white spots that represent concretions. At the base is a conglomerate, 5 feet or more thick, composed of hard, well-rounded pebbles and boulders of the underlying Huronian. The contact is also marked by some brown iron ore. The surface of contact is much like the present coast.

PRE-HURONIAN UNCONFORMITY

Up to this point the discussion has dealt exclusively with the Huronian and later formations. One now turns to a consideration of the ancient crystalline basement or floor upon which these Huronian and later rocks repose. In order, first, to obtain some idea of how enormously much older this crystalline floor is than the Huronian, brief description of the erosion surface is needed.

Sixty years ago, when Logan found that the Huronian sediments rest upon gneiss, he had little means of measuring the time represented by this unconformity, for it must be remembered that he considered all gneisses sediments. Between 1885 and 1890, it became currently known, as a result of Lawson's discovery in western Ontario, that great areas of granite and gneiss are really igneous rocks intrusive in the surrounding formations. Slowly, it came to be the general opinion that these great bodies of granite and gneiss, called batholiths, were found only in mountainous regions and were related to the processes of mountain-building. They were probably the cores of mountain systems. It is obvious that mountains could not be worn down so as to expose hundreds and thousands of square miles of these deep-seated batholiths except during a stupendously long time.

Such was knowledge of the great unconformities that exposed flat floors of batholithic and related rocks when the Cobalt region came under exploration. The pre-Huronian floor was found to be remarkably flat and remarkably undecomposed. It consisted about equally of large areas of granite gneiss and of other areas composed of a great variety of volcanic flows, ash-beds, tuffs, dikes, and other intrusive bodies. There were also some sediments. The volcanics were called Keewatin by Miller, because of their great resemblance to the volcanics of western Ontario. The pre-Huronian was mapped extensively but was not studied with great care during the mining development at Cobalt, because the silver-cobalt ores, with a few prominent exceptions, did not extend into the pre-Huronian. Mineralization was mostly confined to the overlying sediments and diabases.

PRE-HURONIAN SEDIMENTS

The pre-Huronian rocks were simple, consisting mainly of granite gneiss and of older bodies of volcanic rock of many kinds, more or less re-crystallized and digested by the granite. Their chief mineral product was gold, supposing an industry chiefly located around Lake of the Woods and Rainy Lake. The industry languished because the cost of production exceeded the value of the gold produced. But in 1910, some prospectors in northeastern Ontario discovered bodies of quartz carrying good

values of gold. These have become the Dome, Hollinger, McIntyre, and other great mines of the Porcupine or Timmins camp. The discovery of Porcupine was the beginning of a prospecting exploration that is still active and that has yielded such productive camps as Kirkland Lake, Noranda, and many lesser ones.

In the course of this exploration and the geological investigations that have accompanied it, the Keewatin volcanics have been found to be accompanied by thousands of bodies of sediments. These sediments range from small ones, a few yards thick, to others, thousands of feet thick and many miles long. Most of them simply represent interludes of sedimentation in the dominant process of vulcanism. They consist only of Keewatin fragments that have been re-assorted by streams and lakes on the Keewatin surface, for there is abundant evidence for believing that the Keewatin was a land area. Such deposits are devoid of granite pebbles.

The few large sedimentary belts are distinguished by the presence of granite pebbles and boulders in greater or less abundance. Miller found the first of these granite-bearing sediments at the northwest corner of Lake Timiskaming. It is much more highly folded than any of the neighboring Huronian sediments, but its relation to the Keewatin could not be learned. In 1915, a much larger formation, consisting of conglomerate, greywacke, and quartzite, was discovered in the Porcupine mining area. It stands on edge, so it is evidently in great unconformity with any Huronian sediments, which, although absent at Porcupine, lie as nearly flat patches in the country to the south. An angular unconformity of 15 to 20 degrees has been found between the sediments and Keewatin lava flows. Apparently, therefore, there is a major unconformity between the Timiskaming and the Huronian. There is also a major unconformity between the Timiskaming series and the underlying Keewatin. Two such great unconformities would make admirable boundaries for a major time division containing the Timiskaming series and any other formations of like age. Miller was quick to see this. He actually defined such a division, which he called the Timiskamian. It became the practice of the Ontario Department of Mines, thereafter, to include in the Timiskamian any considerable belt of sediments that contained granite boulders and pebbles.

Examination of each of these sedimentary belts does not bear out Miller's theory of a Timiskamian system. For example, in the same year that he found the first Timiskaming series near Haileybury, Coleman discovered at Sudbury a large deposit of sediments, which he placed in the pre-Huronian. This series consists mainly of fine-grained, argillitic material, with evidence of seasonal deposition. It comprises one

principal belt, more than 100 miles long and as much as 8 miles wide; and there are two lesser belts to the south.

The most remarkable thing about this series is that it appears to be continuous upward from the Keewatin. Numerous sections from Keewatin to Sudbury series are all alike. The Keewatin, at the base, contains a few thin deposits of well-stratified, argillitic material. Farther up, these sedimentary deposits become thicker and more numerous, until it becomes a matter of opinion as to whether the rock should be called Keewatin or Sudbury series. This transition continues until there is no Keewatin left. The gradation forms a zone, from a few hundred feet thick to one of greater thickness and of vague boundaries.

This gradation appears to represent a gradual diminution of Keewatin vulcanism and a corresponding increase of sedimentation. It does not indicate any interval of time between Keewatin and Sudbury series.

The Sudbury series, although as extensive a deposit as the Timiskaming, cannot be placed in the Timiskamian because it has no great lower unconformity with the Keewatin.

There is, at Michipicoten, another large deposit of pre-Huronian sediment, called the Doré series. Like the Timiskaming series, it contains an abundance of conglomerate, carrying granite boulders and pebbles. But the Doré series appears to be part of a monocline and to be overlain by many thousands of feet of Keewatin volcanics. Consequently, the Doré series cannot be included in the Timiskamian system of Miller, which is unconformably above the Keewatin.

Again, there is a large belt of sediments in what is known as the Woman River area. These sediments are known as the Ridout series. They have been studied for the Geological Survey by Emmons and Thomson and also by the Ontario Department of Mines. There is some difference of opinion about the structure. Emmons and Thomson considered it to be monoclinal and to be underlain and overlain by large volumes of Keewatin. Like the Michipicoten, the Ridout series could not be included in the Timiskamian system.

The Timiskamian system is, in fact, purely hypothetical. When first described, its boundaries, upper and lower, are unknown. All such pre-Huronian sedimentary belts indicate, by their granite boulders, the existence of some ancient sources for the granite, possibly batholithic. These granites may form the base of a system like the hypothetical Timiskamian, but it is not possible in the present state of knowledge to define such a system. Sedimentary series like the Sudbury series, Doré, Ridout, and others, are probably all local sedimentary deposits on a Keewatin land surface, formed at different times and horizons and not connected

with one another in any one widespread, period of deposition, such as is predicated by their correlation with the Timiskaming series.

ALGOMAN, HAILEYBURIAN, AND MATACHEWAN

When Miller conceived of a Timiskamian system, he found it necessary to assume the existence of two great batholithic invasions, one between the Timiskaming and the Huronian (Cobalt series), the other between the Timiskaming and the Keewatin. Up to this time, all batholithic rocks in northeastern Ontario had been grouped together under the name Laurentian or pre-Huronian. Miller transferred the term Laurentian to those granitic rocks older than his Timiskamian. They consisted solely of granite boulders and pebbles in the Timiskamian sediments. No masses of granitic rocks in place had ever been found, with the possible exception of two small areas in Michipicoten area. For the batholithic rocks that intrude the Timiskaming series, but upon which the Huronian (Cobalt series) lies unconformably and almost flat, he imported the term Algoman from western Ontario. There, Lawson had found batholithic intrusives of two widely different ages, which he called Laurentian and Algoman. The transfer of the younger term to northeastern Ontario seems to have been made solely because both regions contain two granites.

In the course of detailed geological mapping around mining camps, dikes and other small igneous bodies were discovered to intrude the Timiskaming series but to be older than the Algoman granite. In 1920, Miller proposed for these rocks the age term Haileyburian.

In 1923, Miller discovered similar dikes and other small igneous bodies that intrude the Algoman, but are older than the Huronian.

The existence of a Laurentian batholithic invasion is founded on the occurrence of boulders of granitic rocks in the Timiskaming series and other pre-Huronian sedimentary series and on two small masses at Michipicoten, which have not been certainly identified and whose areas are unknown. It follows, therefore, that up to the present time there is no practical need for the name Laurentian. It is also quite uncertain whether the other batholithic rocks of northeastern Ontario are equivalent to Lawson's Algoman or to his Laurentian. Consequently, there is no need in northeastern Ontario for any special name or names for these intrusives: a simple, self-explanatory name, such as pre-Huronian, is sufficient designation.

As the various pre-Huronian sediments, such as the Timiskaming, Doré, and Sudbury series, are not of the same age, it follows that the term Haileyburian can have no time value. It is simply a convenient,

common term for the comparatively few igneous bodies that lie, in time, between a pre-Huronian sediment and the Algoman.

The Matachewan, which applies to rocks between the Algoman and the Huronian, has a somewhat more definite age value. However, as such rocks are not abundant and can only be recognized by detailed geological mapping, this term also is of only local usefulness. It would appear, therefore, that the three names, Algoman, Haileyburian, and Matachewan, have no widespread value or need.

GEOLOGICAL HISTORY IN ORDER OF TIME

In the foregoing part of this paper, the essential events have been described. As nearly as possible, this has been done in order of time, but a good many of the events could not be arranged in an orderly, chronological manner.

In the present section, it is proposed to restate this history in the sequence in which it occurred. Naturally, the events are dealt with in abbreviated form to avoid repetition. The arrangement also takes the form of a geological table, freely annotated. We begin with the oldest division and proceed towards the youngest.

Pre-Huronian—Foundation of the pre-Cambrian Shield

Keewatin

Complex of terrestrial volcanic rocks; much folded, faulted, recrystallized. Profitable gold deposits largely restricted to this division.

Sediments

Many thousands of sedimentary formations largely derived from the Keewatin volcanics; a few mappable as geological units.

Do not form single horizon of sedimentation but occupy different levels in the Keewatin.

Contain granite boulders suggesting invasions during or before the Keewatin.

Batholithic intrusions

Granitic; accompaniments of mountain building which deformed volcanics and sediments.

Source of productive gold deposits in the Keewatin schists.

Major unconformity.

Predominant feature in the geologic history of the Timiskaming sub-province.

	Thickness (Feet)
Huronian—System of sediments and igneous intrusives and extrusives.	
Top not known. Possibly contains another major unconformity between Killarnean intrusion and deposition of White Water series.	
Bruce series—deposited on land or under shallow-water conditions.	
Earthquake movements indicated in Espanola formation.....	5400±
Ramsay Lake conglomerate	
Unassorted, mostly coarse, boulder conglomerate.....	0– 200

	Thickness (Feet)
Mississagi quartzite	
Whitish, feldspathic quartzite, thick-bedded, cross-bedded, few argillaceous members, and some conglomerate.....	0–8000
Bruce conglomerate	
Dark-gray boulder conglomerate; thins toward west, probably disappears west of Sault Ste. Marie.....	20– 800
Espanola formation	
Calcareous siltstone; highly variable proportions of silt and magnesium or iron-bearing carbonate. Dull gray, obscurely banded; ripple marked and sun-cracked. Locally conglomeratic. Thin bed of iron ore at base of two known places	250– 400
Serpent quartzite	
Feldspathic quartzite	
Lower portion resembles Espanola, and is characterized by delicate lamination visible only on weathered surfaces	0–several thousand
Unconformity—time interval probably not great.	
Contact ranges from sharp to gradational in 15-20 feet. Cobalt rests upon an eroded surface ranging from Mississagi to top of Bruce series.	
Cobalt series	6000±
Gowganda formation; glacial deposit. (Logan and Murray's upper slate conglomerate.)	
Boulder clay, varved clays, striated boulders.....	0–4000
Lorrain quartzite	
Massive, green or red, impure at base, to very pure white variety at top. (Upper part forms Lacloche hills along coast of Lake Huron.)	
Quarried at Killarney for metallurgical purposes at Niagara Falls and elsewhere along the lower lakes. Small beds in middle of formation composed entirely of quartz pebbles, which are red, white, or gray. At Bruce Mines, brilliant red pebbles have given the name red jasper conglomerate	0–6000
Banded cherty quartzite	
Slightly calcareous siltstone or fine quartzite deposited in thin, brown to yellow layers; 2000 feet of white and cherty quartzite overlie this.....	725
Nipissing diabase	
Dikes and sills of quartz diabase extending over northeastern Ontario; 5000 square miles of these rocks produced silver-cobalt camp of Cobalt and subsidiaries. One variety contains veins composed of free gold in quartz and breunnerite. Copper produced from diabase in western part of region.	
Killarnean granites	
Batholithic and smaller intrusions, rich in alkalies (soda to potash = 3:2); evidently an accompaniment of Huronian mountain building. Age uncertain; intrude Nipis-	

	Thickness (Feet)
sing diabase; may intrude Sudbury nickel irruptive and if not, a great interval of erosion is implied.	
Whitewater series—Volcanics and sedimentary rocks that fill basin of Sudbury nickel irruptive	
Onaping tuffs	
Coarse tuffs, ranging upward into slate	4500–5000
Onwatin slate	
Black, carbonaceous; grading upward into sandstone	2500
Chelmsford sandstone	
Sooty-colored feldspathic sandstone with small percentage of carbon	80–1500
Nickel irruptive	
Saucer-shaped sill or laccolith, 37 by 17 miles.	
Differentiated from norite to micropegmatitic granite.	
Source of Sudbury nickel-copper deposits.	
Trap and olivine diabase dikes.	
Keweenawan flows and sediments—Several small areas on east coast of Lake Superior represent outliers of Keweenawan.	
Major unconformity	
Paleozoic sediments	
Chiefly Cambrian and Ordovician	
Major unconformity	
Glacial and Recent sediments	

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THEORY OF SCALE MODELS AS APPLIED TO THE STUDY OF GEOLOGIC STRUCTURES

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INTRODUCTION

Many of the phenomena of physical science are simple enough and well enough understood that they are amenable to complete mathematical analysis without recourse to auxiliary experimentation. There are other phenomena, however, which, though being made up of well-understood simple systems, are so complicated as a whole as to render complete mathematical analysis difficult or impossible. The distribution of stress in a complicated machine part, or the flow of water in an irregularly shaped vessel, would constitute examples of the latter kind.

When something must be known about one of these more complicated problems it is usual, whenever possible, to obtain the desired information empirically by direct experimentation. Often, however, the thing

studied is too large to be experimented with. Or, as in the case of large engineering structures, the information on a bridge, dam, or building is needed in advance of designing the structure.

Under these conditions, where mathematical analysis is inadequate, and where for one reason or another direct experimentation is precluded, the best remaining alternative is to construct and study a scale model. This in fact is what is being done successfully in aerodynamical, hydraulic, and mechanical and electrical engineering. Aerodynamical studies are being made on scale models of aeroplanes in wind tunnels. In modern hydraulic laboratories such as that of the United States Army Engineering Corps at Vicksburg, Mississippi, harbors, canals, rivers, ground water flow, and other similar problems are being studied by means of models. Civil engineers are studying models of buildings, bridges, and similar structures. Mechanical engineers are making transparent models of machine parts and studying their stress distributions by the strain figures revealed by crossed nicol prisms and polarized light. Electrical engineers are studying both electrical models of electrical systems, and what may seem surprising, electrical models of mechanical systems.

The geological problems of mountain making and of diastrophism in general are peculiarly of the type that do not lend themselves readily to analysis, and the size of the elements involved places them beyond the range of direct experimentation. In this case also there remains the alternative of studying such phenomena by means of experiments performed upon properly built small scale models. For more than a hundred years attempts have been made to study the mechanics of the rock deformation in mountain making by this means. Among the earliest of such experiments were those of Sir James Hall [1] who studied the formation of folds by the use of models, employing layers of cloth in some instances, and of clay in others, to represent strata. A. Daubree [2] was a somewhat later pioneer. He performed a wide range of experiments studying both fracture and folding. For materials, he used glass, plaster, wax, and strips of metal. Later came Willis' [3] well-known experiments on the mechanics of the Appalachian type of structure. Willis used a pressure box allowing for a model about a meter in length. The materials used in his various models were composed of layers of plaster of paris, and of waxes of various consistencies, weighted from above by a heavy load of shot.

Subsequently, numerous others have performed model experiments to elucidate various geological problems. A few of the more recent ex-

[1] W. Paulcke: *Das Experiment in Geologie* (1912) Karlsruhe.

[2] A. Daubree: *Etudes synthétiques de géologie expérimentale*, pt. 1 (1879) Paris.

[3] Bailey Willis: *The mechanics of Appalachian structure*, U. S. Geol. Surv., 13th Ann. Rept., pt. 2 (1891-1892) p. 211-289.

perimenters have been Chamberlin and Miller,[4] Mead,[5] Link,[6] Hans Cloos,[7] Fujiwhara, Tsujimura and Kusamitsu,[8] Ph. Kuenen,[9] and Nettleton.[10] Most of these experimenters employed materials ranging in strength from plaster to rather soft waxes and clay. Cloos, Kuenen, and Nettleton differed from the others by using materials of extreme weakness. In his experiments on tectonic structures, Cloos used an almost liquid clay. Kuenen used china clay, vaseline, mineral oil, and paraffin at a temperature only slightly below that of melting. He tried gypsum but discarded it as being too strong. Nettleton, in experiments representing salt domes, employed two viscous liquids of different densities.

Much excellent experimentation has also been done by Adams[11] and associates, von Kármán,[12] Bridgman,[13] Griggs,[14] and others, on the elastic and plastic properties of rocks. This kind of work provides some of the necessary data for model experimentation but it is not to be classed as model experimentation itself.

Ever since the days of John Hutton[15] there have been some geologists who have maintained a legitimate skepticism regarding the significance of the results obtained by means of models of plaster and clay, when applied to mountain structures where the material is hard rock. Besides the discrepancy in the strengths of the materials in the two cases, the. has also been the disparity in the time required, the model being de-

[4] R. T. Chamberlin and W. Z. Miller: *Low-angle faulting*, Jour. Geol., vol. 26 (1918) p. 1-44.

[5] Warren J. Mead: *Notes on the mechanics of geologic structures*, Jour. Geol., vol. 28 (1920) p. 505-523.

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[10] L. L. Nettleton: Fluid mechanics of salt domes, Am. Assoc. Petrol. Geol., Bull. vol. 18 (1934) p. 1175-1204.

[11] F. D. Adams and J. T. Nicholson: *An experimental investigation into the flow of marble*, Royal Soc. London, Philos. Tr., ser. A, vol. 195 (1901) p. 363-401.

F. D. Adams: *An experimental investigation into the action of differential pressure on certain minerals and rocks, employing the process suggested by Professor Kick*, Jour. Geol., vol. 18 (1910) p. 489-525.

[12] Th. von Kármán: *Festigkeitversuche unter allseitigen druck*, Zeitschr. des Vereins deutscher Ingeniere, vol. 55 (1911) p. 1749-1757.

[13] P. W. Bridgman: *The physics of high pressure* (1931) (New York); *Shearing phenomena at high pressure of possible importance to geology*, Jour. Geol., vol. 44 (1936) p. 653-669.

[14] David T. Griggs: *Deformation of rocks under high confining pressures*, Jour. Geol., vol. 44 (1936) p. 541-577.

[15] *See* Sir Archibald Geikie: *The founders of geology* (1897) London. Also W. Paulcke: *Das experiment in Geologie* (1912) Karlsruhe.

formed in a few hours by an amount which in the mountain may have required a million years.

Still another reason for questioning the validity of model experiments is to be found in the artifices adopted by some of the experimenters. Willis, for example, found it necessary to load his models of Appalachian structures with an overburden of 3 to 5 pounds per square inch of lead shot to make them stay down properly. This, on the scale of his models, was equivalent to an overburden of something like 100 miles of sediments. Even then the more competent plaster layers broke into rigid slabs instead of folding plastically, as even the quartzites in the Appalachians themselves are known to have done.

In view of the diversity of the materials that have been used in model experiments, and of the wide range of methods employed by various experimenters, one wonders which, if any, of the results are trustworthy. Particularly there is the need for an objective criterion to enable one to determine what the correct properties of a model should be for the best similarity, when the properties of the original are known, or whether it is even possible to build a correct model from available materials.

Dating from the time of Galileo [16] there has been accumulating a body of knowledge of how a model is related to its original, or of how the physical properties of a body change with change of size. During the last 50 years this body of knowledge has been brought to a high state of advancement in the various branches of engineering, especially in connection with the work in modern hydro- and aerodynamic laboratory practice using scale models.[17] Recently Churchill [18] has worked out the theory for problems involving the conduction of heat.

It is the purpose of the present paper to derive the general theory of the similarity between a model and its original for purely mechanical systems. This theory will then be applied to a number of illustrative geological problems and a set of model ratios between properties of the model and the corresponding properties in the original will be determined. This will enable us, if we know given properties of the original, to determine what the corresponding properties of the model should be. The knowledge thus acquired may be used either as a guide in actually performing experiments or as a criterion for evaluating experiments which have already been performed. In many cases it will allow one

[16] Galileo Galilei: *Dialogues concerning two new sciences* [translation by Henry Crew and Alfonso de Salvio] (1914) p. 109-152. New York.

[17] Francesco Marsola: *Some considerations regarding hydraulic models*, Hydraulic Laboratory Practice, 1926-1929, A.S.M.E. (1929) appendix 13, p. 743-758.

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L. Prandtl and O. G. Tietjens: *Applied hydro- and aeromechanics* (1934) Chap. 2. New York.

[18] R. V. Churchill: *Comparison of the temperatures in a solid and its scaled model*, Physics, vol. 6 (1935) p. 100.

to bring a problem of which he has little sensory experience down to a scale of things with which he is already familiar, and thereby enable him to have some intuitive understanding of it, even if no experiment is performed.

It is possible to extend dimensional analysis (which is what we shall be using) to include thermodynamic relations accompanying mechanical changes. To attempt to do this here, however, would complicate our problem more than the results to be obtained seem to warrant. Consequently we shall confine our attention chiefly to the mechanical aspects of the problems considered.

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The present paper is in a sense the integration of many of the writer's past experiences, by no means the least important of which have been conversations with friends. Among those friends who perhaps unknowingly have contributed most along these lines have been J Harlen Bretz, J. T. Stark, Waldo W. Waring, H. W. Straley III, Robert Thomson, Robert E. Bates, Girard E. Wheeler, Norman H. Dolloff, and Howard Scott.

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THEORY OF MECHANICAL SCALE MODELS

FAMILIAR EXAMPLES

Before going into the intricacies of the theory of models it might be pointed out that there are a large number of familiar, everyday occurrences that can be shown to be the direct consequence of the dimensional relations we are to derive and use.

1. Of two animals more or less geometrically similar but of different sizes, the smaller can fall without injury through a greater height, measured in multiples of its own length, than the larger. A mouse can fall without injury a distance many times his own length, a dog fewer, and a horse fewer still.

2. Of two animals of different sizes but similar shapes, the bodily motions, (leg motions, voice frequency, heart beat, etc.) of the smaller have the higher frequency.

3. A windmill of small radius will turn faster than one of large radius in the same wind.

4. Given any two machines having the same rate of output of the same product, the one whose parts have the higher frequency will, in general, be the smaller. For example, a high speed shaft need be only a fraction the size of a low speed shaft for identical transmission of power. This

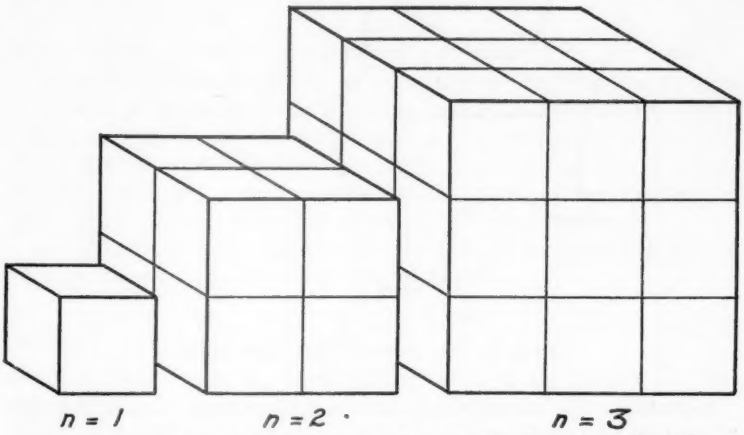

FIGURE 1.—*Change of magnitudes of a cube as the length of side is changed*

is one of the most important facts confronting the mineral industries today.

5. In electrical equipment, for the same power, the higher the frequency the smaller the transformers, generators, condensers, and motors. Short wave radio parts are very much smaller than those used for broadcast frequencies, which are smaller than audio-frequency parts, and all are smaller than 60-cycle power equipment of the same power.

Although the reasons for the above relations may not be obvious, they serve to demonstrate the fact that when the size of a body is changed its various other physical properties also change, but, in general, not proportionally to the change in size.

It shall be our immediate task to investigate how various physical properties of the body change as its size is changed. Consider a cube of lead, for example. Now let us enlarge it n diameters without changing its density. Let n be successively 1, 2, 3, 4, 5, and so on. Let us then investigate the manner in which the area, the volume, the mass, the weight, and the pressure at the base of the cube vary for the different values of n.

When n is 1, we will consider each of these quantities as unity—unit length of side, unit mass, unit area, unit volume, unit weight, and unit

pressure. Now let n be 2. The length of side will be 2, the area 4, the volume 8, the mass 8, the weight 8, and the pressure, which is the weight divided by the area of the base, 2.

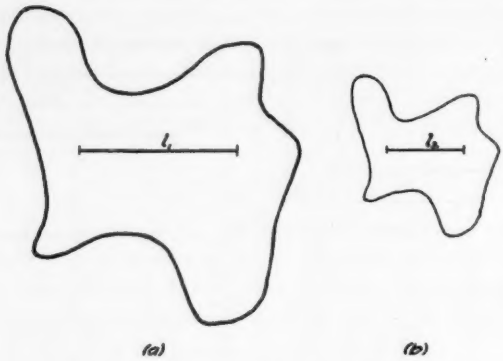

FIGURE 2.—*Two geometrically similar bodies*

Let n be 3. The side will be 3, the area 9, the volume, mass, and weight 27, and the pressure 3.

This is shown in Table 1 for values of n up to 5.

TABLE 1.—*Changes in physical properties with change in size of body*

n	Length	Area	Volume	Mass	Weight	Pressure
1	1	1	1	1	1	1
2	2	4	8	8	8	2
3	3	9	27	27	27	3
4	4	16	64	64	64	4
5	5	25	125	125	125	5

From this it is clear that the area increases as the square, the volume, mass, and weight as the cube, and the pressure at the base as the first power of the enlargement factor of the linear dimension of a cube, gravity and the density remaining constant.

DEGREES OF SIMILARITY

Geometrical Similarity.—Two bodies are said to be geometrically similar when all corresponding lengths are proportional and all corresponding angles of the two bodies are equal. This is simply a generalization

of the familiar definition of similar triangles from geometry, extended to include bodies of any shape.

If we let l_1 be a length in one of the bodies, and l_2 the corresponding length in the other, then

$$\frac{l_2}{l_1} = \lambda, \text{ or } l_2 = \lambda l_1, \tag{1}$$

where λ is the *constant of proportionality* of length of the two bodies.

If the first body is thought of as the *original* and the second as a replica of it made to scale, the latter is said to be a *scale model* of the former. The model may be either a reduction or an enlargement of the original; if a reduction, λ will be less than unity, if an enlargement, λ will be greater than unity. λ is the model ratio of length. It will be observed that λ is the familiar map scale extended to three dimensions.

In two geometrically similar bodies the ratio of corresponding areas is equal to the square of the model ratio of length.

$$\frac{A_2}{A_1} = \lambda^2, \tag{2}$$

where A_2 and A_1 are the corresponding areas of the second and first respectively. This can be seen at once if the two areas are divided into similar grid-works of n squares each. Then for each square

$$\frac{l_2^2}{l_1^2} = \lambda^2,$$

and for n such squares

$$\frac{A_2}{A_1} = \frac{nl_2^2}{nl_1^2} = \lambda^2.$$

In two geometrically similar bodies the ratio of corresponding volumes is

$$\frac{V_2}{V_1} = \frac{nl_2^3}{nl_1^3} = \lambda^3. \tag{3}$$

Kinematic Similarity.—If two geometrically similar bodies undergo geometrically similar changes of shape or positions, or both, the two bodies are said to be kinematically similar provided the time required for any given change in the one is proportional to that required for the corresponding change in the other.

$$\frac{t_2}{t_1} = \tau, \tag{4}$$

where t_1 is the time required for the original to undergo a given transformation, t_2 the corresponding time required for the model, and τ the *model ratio of time.*

Similarity of Velocities and Accelerations.—If two bodies are kinematically similar the velocities and accelerations of corresponding points must be proportional:

$$\frac{v_2}{v_1} = \eta = \frac{\frac{l_2}{t_2}}{\frac{l_1}{t_1}} = \frac{\lambda}{\tau} = \lambda\tau^{-1}, \tag{5}$$

where v_1 and v_2 are the velocities of corresponding points and η the *model ratio of corresponding velocities.*

For accelerations:

$$\frac{a_2}{a_1} = \gamma = \frac{\frac{l_2}{t_2^2}}{\frac{l_1}{t_1^2}} = \lambda\tau^{-2}, \tag{6}$$

where γ is the *model ratio of acceleration.*

For angular velocities:

$$\frac{\omega_2}{\omega_1} = \frac{\frac{\theta_2}{t_2}}{\frac{\theta_1}{t_1}} = \frac{\theta_2}{\theta_1} \cdot \frac{t_1}{t_2}, \tag{7}$$

where ω_1 and ω_2 are the angular velocities, and θ_1 and θ_2 the angles of rotation in the times t_1 and t_2 in the original and the model respectively.

But

$$\theta_1 = \theta_2.$$

Therefore

$$\frac{\omega_2}{\omega_1} = \frac{1}{\tau} = \tau^{-1}. \tag{8}$$

Dynamic Similarity.—In the discussion of geometrical and kinematic similarity we have described the relations between the forms and motions of bodies without regard to the fact that they possess mass. As all bodies possess mass, however, the presence of a gravitational field and the inertial reactions of mass to accelerated motion set up forces which have to be reckoned with.

To begin with, we require that our model have a mass distribution similar to that of the original. What we mean by this is that if dm_1 be the mass of an element of volume, dV_1, in the original, and dm_2 that of the corresponding element of volume, dV_2, of the model, then the ratio

$$\frac{dm_2}{dm_1} = \mu, \tag{9}$$

must hold point by point throughout the two bodies, μ being the *model ratio of mass.*

From the model ratios of mass and length, the ratio of density is obtained directly. Density is defined as the mass divided by the volume. Hence the model ratio of density δ is given by

$$\delta = \frac{\rho_2}{\rho_1} = \frac{\dfrac{dm_2}{dV_2}}{\dfrac{dm_1}{dV_1}} = \mu\lambda^{-3} \tag{10}$$

and is a constant throughout the two bodies.

The forces acting upon any element of mass, dm, occupying an element of volume, dV, may be divided into two separate classes, the *body forces* and the *surface forces.* The body forces, as the name implies, are forces originating inside the body of the volume. In purely mechanical systems the body forces are of two kinds, those due to gravity and those due to inertia. If we call these respectively f_g and f_i, we then have

$$f_g = dm \cdot g \tag{11}$$

and by Newton's Second Law of Motion

$$f_i = dm \cdot a \tag{12}$$

where g and a are the forces per unit of mass due to gravity, and to the local acceleration of the mass, respectively.

The surface forces are forces acting only on the external surface of the element of volume considered. They are consequently proportional to the magnitude of the area acted upon. The intensity of surface forces is measured by the ratio of the force to the area acted upon, and is known as the *stress.*

Hence,

$$\text{Stress} = \frac{\text{Force}}{\text{Area}},$$

or is the force acting per unit of area. One of the most familiar examples of stress is *pressure,* where the force is directed toward the area acted upon and normal to it.

Pressure is, however, only a special case of stress. Another special case is *tension,* where the force is directed away from the surface acted upon, and normal to it. The most general type of stress, however, is that in which the force is oblique to the surface. In this case it may be resolved into a component normal to, and another parallel to, the surface acted upon. The former per unit of area is known as the *normal component of stress,* and the latter as the *tangential* or *shearing component.*

The resultant surface force acting over any volume V is the sum obtained by adding vectorially the products of the stresses by the areas acted upon over the entire surface.

Surface forces originate in several ways. In the case of fluids at rest, they consist entirely of pressures. The shearing component is non-

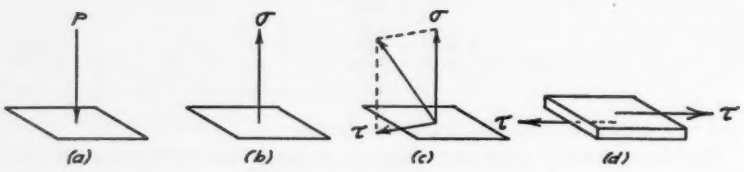

FIGURE 3.—*Types of stress*

(*a*) Pressure; (*b*) tensile stress; (*c*) general stress, showing normal and tangential components; (*d*) shearing stress.

existent. In the case of elastic solids, surface forces result from both normal and shearing components of stress related to the elastic strain. In the case of fluids and plastic bodies in motion the surface forces re-

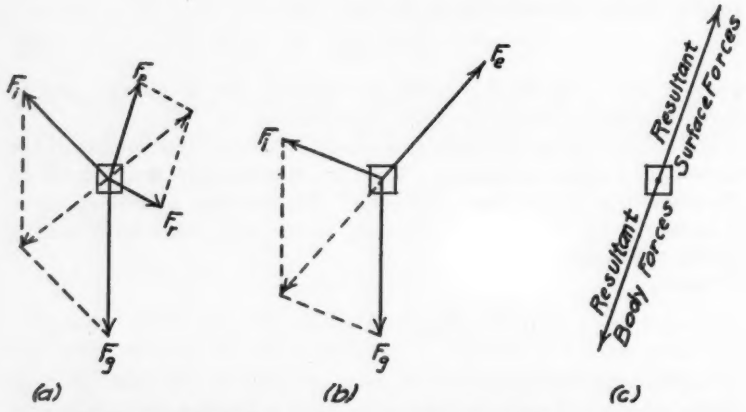

FIGURE 4.—*Equilibrium of forces acting upon an element of mass*

(*a*) in viscous body, (*b*) in elastic body, (*c*) general equilibrium between resultant body forces and resultant surface forces.

sult from, or are related to, the volume change and the time rate at which the body undergoes deformation. Those resulting from the latter cause may be thought of as the *resistive* forces.

We now have body forces due to gravity and to inertial resistance to acceleration and surface forces arising from pressure, from elastic

strains of the body, and from viscous resistance to rate of change of form. In general, each element of mass is in a state of dynamic equilibrium among as many of these forces as are acting simultaneously, and we may say that the vector sum of the body forces acting is equal and opposite to the vector sum of the surface forces.

Let F_g, F_p, F_r, F_e, and F_i be respectively the resultant vectors of forces due to gravity, to pressure, to viscous resistance, to elasticity, and to inertia, acting upon an element of mass dm contained in volume

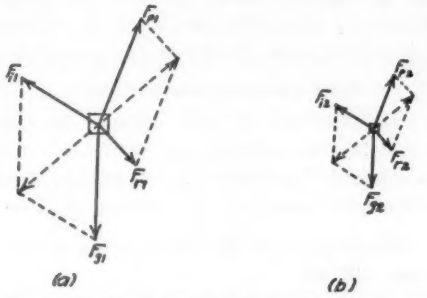

FIGURE 5.—*Similarity of forces acting on corresponding elements of mass in viscous bodies*
(*a*) Original, (*b*) model.

dV. As these are all the forces acting, their vector sum must be zero, so we have for bodies undergoing fluid or plastic deformation

$$F_g + F_p + F_r = -F_i, \tag{13}$$

and for elastic solids

$$F_g + F_e = -F_i. \tag{14}$$

Dynamic similarity requires that on each element of mass dm_2 occupying volume dV_2 of the model the forces acting must be such that the motion is geometrically and kinematically similar to the corresponding motion of the element of mass dm_1 in volume dV_1 of the original. This condition is satisfied exactly if for each force F_1 acting on mass dm_1 in the original there is a corresponding vector force F_2 acting upon mass dm_2 in the model, F_2 having the same orientation as F_1, and the ratio of the magnitudes of the two forces being

$$\frac{F_2}{F_1} = \phi \tag{15}$$

where ϕ is the *model ratio of force* and is constant for all corresponding forces throughout the two bodies.

We have accordingly

$$\frac{F_{g2}}{F_{g1}} = \frac{F_{p2}}{F_{p1}} = \frac{F_{e2}}{F_{e1}} = \frac{F_{r2}}{F_{r1}} = \frac{F_{i2}}{F_{i1}} = \phi \tag{16}$$

for all corresponding forces.

We now have as the most general condition for similarity that *two massive bodies can be geometrically and kinematically similar only provided the masses of the one, point by point, are proportional to the corresponding masses of the other, and that the corresponding forces, point by point, have the same directions and proportional magnitudes.*

FORCES DUE TO VARIOUS CAUSES

Statement of Investigation.—It now remains for us to investigate somewhat more in detail the different species of force that may act upon an element of mass within a body, for, as we have seen, the ratio of each of these between the model and the original must be the constant ϕ.

Body Forces.—We begin with the body forces whose magnitudes by equations (11) and (12) are

$$F_g = dm \cdot g,$$

$$F_i = dm \cdot a,$$

and by (16) together with (11) and (12) we obtain

$$\frac{F_{g2}}{F_{g1}} = \frac{dm_2 g_2}{dm_1 g_1} = \phi = \frac{F_{i2}}{F_{i1}} = \frac{dm_2 a_2}{dm_1 a_1} \tag{17}$$

But since

$$\frac{dm_2}{dm_1} = \mu$$

and

$$\frac{a_2}{a_1} = \lambda\tau^{-2}$$

then

$$\phi = \frac{F_{i2}}{F_{i1}} = \mu\lambda\tau^{-2}. \tag{18}$$

By equations (16) we learned that all corresponding forces must have the same ratio ϕ, and now by (18) we learn that ϕ is uniquely determined by the values assigned to μ, λ, and τ, the fundamental model ratios of mass, length, and time, which, in the most general case, may be chosen independently and arbitrarily.

The general case is merely a statement of the fact that where both gravitative and inertial forces coexist, both must have the same model

ratio ϕ. This fact, however, constitutes a certain duality of control in that the ratio of inertial forces must be made to fit the ratio of the gravitative forces, which lessens considerably the degree of arbitrariness that otherwise would exist. In particular, when both the original and the model are on the earth's surface

$$\frac{g_2}{g_1} = \gamma_g = 1 = \gamma = \lambda\tau^{-2}, \tag{19}$$

in which case

$$\tau = \lambda^{\frac{1}{2}}. \tag{20}$$

Consequently under such restraint λ and τ are no longer mutually independent, for when either is arbitrarily chosen the other is uniquely determined; only μ and λ, or μ and τ remain as completely independent model ratios.

In separate spherical astronomical bodies, however,

$$g = K \cdot \frac{m}{r^2}, \tag{21}$$

where g is the attractive force on a unit mass at any point outside the

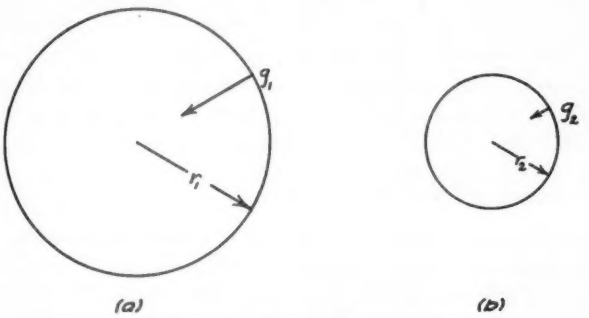

FIGURE 6.—*Similarity of gravitational forces*

body at distance r from its center, and K the universal constant of gravitation. Then if we had two such bodies, one a model of the other, we should have for the ratios,

$$\frac{g_2}{g_1} = \frac{K_2 \cdot \dfrac{m_2}{r_2^2}}{K_1 \cdot \dfrac{m_1}{r_1^2}} = \gamma = \lambda\tau^{-2},$$

or

$$\frac{K_2}{K_1} = \gamma\mu^{-1}\lambda^2 = \mu^{-1}\lambda^2\tau^{-2}. \tag{22}$$

While it is not physically possible to alter the constant of gravitation it is sometimes useful for theoretical purposes (as we shall see later) to be able to compute a model as if K could be varied. Equation (22) enables us to do this in terms of fundamental model ratios μ, λ, and τ.

Surface Forces.—We turn now to the surface forces which, as we have seen, must be such as to balance the body forces. If we take a small element of volume dV containing mass dm, the resultant surface forces over the element must be equal and opposite to the body forces acting on the mass dm. The surface forces are proportional to the stress and the area; the body forces are proportional to the mass, and hence to the volume. If l is a characteristic length of the element of volume, we have as the ratio of the body forces to the surface forces,

$$\frac{\text{Body Forces}}{\text{Surface Forces}} = k\cdot\frac{l^3}{l^2}, \tag{23}$$

which tends to the limit zero as l tends to zero. In other words the body forces diminish with size more rapidly than the surface forces, so that if the element of volume is small enough we may completely neglect the body forces for that element, and focus our attention upon its surface forces, which must then be in equilibrium among themselves. This leads us to the conception of the state of stress at a given point within the body. The criterion of dynamic similarity then tells us that at every corresponding point within the two bodies the states of stress must be similar at corresponding times.

State of Stress at a Point.—Let us take an infinitesimal element of volume with lengths of sides dx, dy, and dz, and small enough that the body forces are negligible compared with the surface forces. We then have as the condition of equilibrium that all the surface forces must be in equilibrium among themselves; that is, there must be no resultant force tending to accelerate the body, nor a torque tending to rotate it.

We resolve the stresses acting on each face into a normal component and two shearing components parallel to each of the three axes X, Y, and Z. We designate the normal component by the letter σ and we give it a subscript x, y, or z, designating that the face upon which it acts is perpendicular to the X, the Y, or the Z axis respectively. Each shearing stress component is designated by the letter τ (not to be confused with the model ratio of time) bearing two subscripts, the first

signifying the axis perpendicular to the face acted upon, and the second the axis to which the component is parallel.

For each pair of faces there is one face for which the outward directed normal (positive normal stress) is directed toward the positive end of the axis of coordinates to which it is parallel and one for which

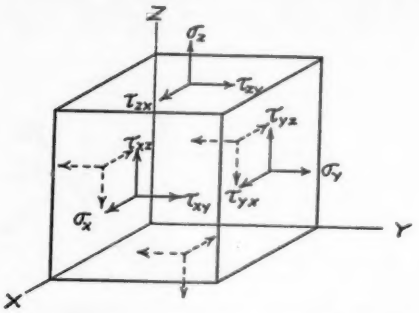

FIGURE 7.—*Stresses acting on an infinitesimal elementary cube*

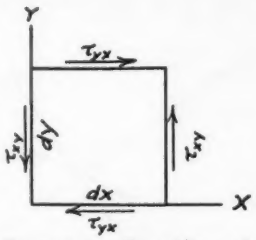

FIGURE 8.—*Equality of shearing stress couples*

Clockwise couple tending to produce rotation about the Z axis is equal and opposite to the counter-clockwise couple.

the normal is oppositely directed. We adopt the convention that on any face for which the outwardly directed normal points toward the positive end of the axis to which it is parallel, the two components of shearing stress are positive when their vectors point toward the positive ends of the axes to which they are parallel. On the opposite faces of each pair all stress directions become reversed. The normal stresses are positive when directed outward.

We then have on the three pairs of faces of the elementary parallelopiped the following components of stress:

$$\begin{array}{lll} \sigma_x & \tau_{xy} \quad \tau_{xz} & \text{on the } X \text{ face,} \\ \sigma_y & \tau_{yz} \quad \tau_{yx} & \text{on the } Y \text{ face,} \\ \sigma_z & \tau_{zx} \quad \tau_{zy} & \text{on the } Z \text{ face.} \end{array}$$

This number can be reduced when we consider that the turning moments about each of the axes must be zero. Then

$$\tau_{xy} = \tau_{yx}; \ \tau_{xz} = \tau_{zx}; \ \tau_{yz} = \tau_{zy}. \tag{24}$$

This leaves us with six fundamental components of stress:

$$\begin{array}{lll} \sigma_x, \ \sigma_y, \ \sigma_z & & \text{normal stresses,} \\ \tau_{xy}, \ \tau_{yz}, \ \tau_{zx} & & \text{shearing stresses.} \end{array}$$

Principal Stresses.—While the proof of this will not be developed here, it can be shown that for any point in a body it is always possible to so

orient the X, Y, and Z axes that all of the components of shearing stress vanish:

$$\tau_{xy} = \tau_{yz} = \tau_{zx} = 0.$$

In such an instance there remain only the normal components of stress, σ_x, σ_y, and σ_z, which are then designated σ_1, σ_2, and σ_3 and are said to be the three *principal stresses* at the point in question. If the

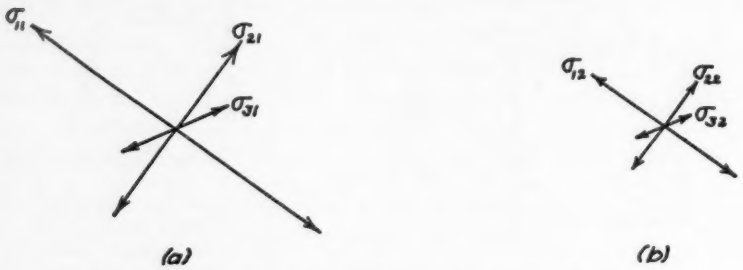

FIGURE 9.—*Similarity of principal stresses at corresponding points* (*a*) in original, (*b*) in model.

magnitudes and orientations of the three principal stresses at any point are known, the components of normal and shearing stress on any plane through that point can be computed.

In the most general case of stress in rigid bodies or in viscous fluids in motion the three principal stresses are unequal. For fluids at rest all three become equal and all shearing stresses in whatever directions become zero.

Strain.—In an elastic body the normal stresses produce elongations in the directions of the three axes, and the shearing stresses produce shearing deformation. All changes of size or of shape of a body, or both, are said to constitute *strains*. Strains are measured in terms of the ratio of the increase in a given geometrical dimension to the original value of that dimension. Thus:

$$\text{Volume strain} = \frac{\text{Increase of volume}}{\text{Original volume}}, \tag{25}$$

$$\text{Length strain} = \frac{\text{Increase of length}}{\text{Original length}}, \tag{26}$$

$$\text{Shear strain} = \frac{\text{Shear displacement}}{\text{Thickness}}. \tag{27}$$

Elasticity.—The relation between strain and stress in elastic bodies is given by Hook's Law which states that for a given material

$$\frac{\text{Stress}}{\text{Strain}} = \text{a constant.} \tag{28}$$

This constant is called the *modulus of elasticity* of the material.

As we have volume strain, length strain, and shear strain, associated with appropriate stresses, there must be as many different kinds of

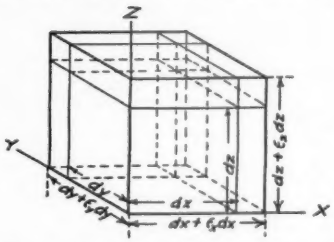

FIGURE 10.—*Elongations parallel to axes of coordinates*

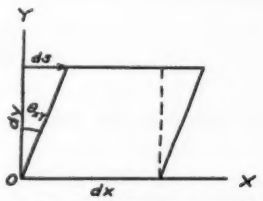

FIGURE 11.—*Shear parallel to X axis only*

moduli of elasticity, namely the *volume modulus, length* or *Young's modulus,* and *shear* or *rigidity modulus.*

In the case of our elementary volume we may designate the length strains parallel to the three axes respectively as ε_x, ε_y, and ε_z where ε is the amount of elongation per unit of original length.

Shear strain, from definition (27), involves a distortion of the kind obtained by shearing a deck of cards, one over the other. This involves a change of the angle at the end of the deck by an amount θ which may be taken as a measure of the strain. More strictly the shear is equal to the tangent of θ but when θ is small this becomes equal to the radian measure of θ itself. Elastic strains seldom are great enough that one need consider the tangent of θ as being different from θ itself.

Now we relate the stresses acting upon our elementary volume to the strains produced. (See Fig. 12.) ε_x, ε_y, and ε_z are the elongations; θ_{xy}, θ_{yz}, and θ_{zx} are the shears parallel to the X-Y, the Y-Z and the Z-X planes respectively. We then have

$$\frac{\sigma_x}{\epsilon_x} = \frac{\sigma_y}{\epsilon_y} = \frac{\sigma_z}{\epsilon_z} = E, \tag{29}$$

where E is the modulus of elongation, or Young's modulus; and

$$\frac{\tau_{xy}}{\theta_{xy}} = \frac{\tau_{yz}}{\theta_{yz}} = \frac{\tau_{zx}}{\theta_{zx}} = G, \tag{30}$$

where G is the modulus of rigidity.

From these it is possible to derive also the volume modulus though there is no necessity of our doing so for the present purposes.

What we are leading up to is the question of what must be done to the moduli of elasticity in order that the model may be dynamically similar to the original. We can now write that for similarity the ratio

$$\frac{E_2}{E_1} = \frac{\dfrac{\sigma_2}{\epsilon_2}}{\dfrac{\sigma_1}{\epsilon_1}} = \frac{\sigma_2}{\sigma_1} \cdot \frac{\epsilon_1}{\epsilon_2}, \tag{31}$$

and that

$$\frac{G_2}{G_1} = \frac{\dfrac{\tau_2}{\theta_2}}{\dfrac{\tau_1}{\theta_1}} = \frac{\tau_2}{\tau_1} \cdot \frac{\theta_1}{\theta_2}, \tag{32}$$

and more generally that

$$\frac{(\text{Modulus})_2}{(\text{Modulus})_1} = \frac{\dfrac{(\text{Stress})_2}{(\text{Strain})_2}}{\dfrac{(\text{Stress})_1}{(\text{Strain})_1}} = \frac{(\text{Stress})_2}{(\text{Stress})_1} \cdot \frac{(\text{Strain})_1}{(\text{Strain})_2} \tag{33}[19]$$

We have already seen that all strains of whatever kind are measured in homogeneous units of length/length, volume/volume, etc. As geometrical similarity requires all corresponding lengths to be proportional and corresponding angles to be equal in the model and the original, it follows that all strains are likewise equal. Hence the model ratio of strain is unity.

Consequently the model ratio of the elastic moduli between the model and the original is equal to that of stress. Each stress is the ratio of a force to an area. We have for the model ratio of stress, and consequently of the moduli of elasticity,

$$\frac{(\text{Modulus})_2}{(\text{Modulus})_1} = \frac{(\text{Stress})_2}{(\text{Stress})_1} = \sigma = \frac{\dfrac{f_2}{A_2}}{\dfrac{f_1}{A_1}} = \mu\lambda^{-1}\tau^{-2}, \tag{34}$$

where σ is now taken as the model ratio of stress and τ that of time, as originally.

Viscosity.—In the case of fluids we obtain the stresses at a point by identically the same reasoning as that employed for the case of elastic bodies and again the six components of stress are:

$$\sigma_x, \quad \sigma_y, \quad \sigma_z,$$
$$\tau_{xy}, \quad \tau_{yz}, \quad \tau_{zx}.$$

[19] Strictly speaking this is true only for isotropic bodies. For anistropic materials the moduli of elasticity are functions of the direction. This, however, is not incompatible with similarity provided that the model be given an anisotropy similar to that of the original.

Since our interest attaches only to changes of form in relatively incompressible materials we may consider the fluids to be incompressible. In this case we may disregard the normal components of stress, which tend to produce change of volume only, and focus our attention upon the shearing components.

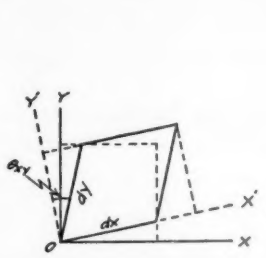

FIGURE 12.—*Shear parallel to both X and Y axes*

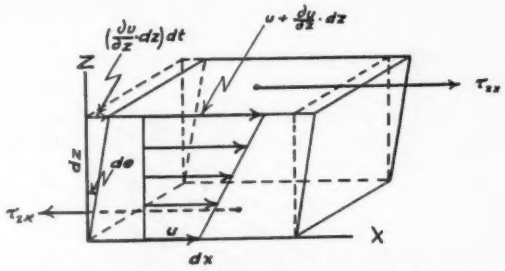

FIGURE 13.—*Relation between rate of shear, velocity gradient, and shearing stress parallel to X axis*

In its behavior under shearing stress a fluid differs from an elastic body in that whereas the shearing stress is proportional to the shearing strain in the latter, in the case of the former the shearing stress is proportional to the time rate at which the shearing strain occurs.

If we let $d\theta_{xy}$, $d\theta_{yz}$, and $d\theta_{zx}$ be the angles of the shear occurring in the fluid in planes parallel, respectively, to the X-Y, the Y-Z, and in the Z-X planes, in the time dt, then we have:

$$\left.\begin{aligned} \tau_{xy} &= \xi \frac{d\theta_{xy}}{dt} \\ \tau_{yz} &= \xi \frac{d\theta_{yz}}{dt} \\ \tau_{zx} &= \xi \frac{d\theta_{zx}}{dt} \end{aligned}\right\} \tag{35}$$

where ξ is the coefficient of viscosity of the fluid.

To determine the model ratio of the coefficients of viscosity, we solve (35) for ξ and take the ratio

$$\frac{\xi_2}{\xi_1} = \frac{\tau_2}{\tau_1} \cdot \frac{d\theta_1}{d\theta_2} \cdot \frac{dt_2}{dt_1} = \psi, \tag{36}$$

or

$$\psi = \sigma\tau = \mu\lambda^{-1}\tau^{-1}, \tag{37}$$

where ψ is the model ratio of viscosities and τ, when used as a model ratio, signifies that of time.

Reynold's Number.—In the case of fluid motion somewhat more light is thrown on the requirements for similarity if we focus our attention upon the ratio of the body force due to inertia acting upon the mass

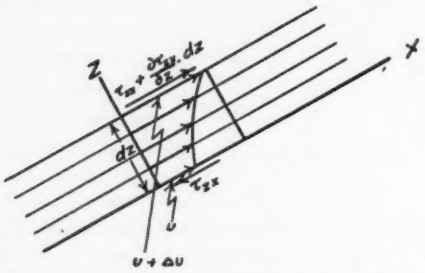

FIGURE 14.—*Shearing stress gradient and velocity gradient*

Gradient of shearing stress will produce a net force in the X direction.

contained in an element of volume of the fluid, to the force due to viscosity acting upon the surface of that volume. In this case we choose our element of volume large enough that the body forces are no longer negligible in comparison with the surface forces.

The force due to inertia is equal to the mass multiplied by the acceleration and is directed opposite to that of the acceleration. The net force due to viscous resistance of the fluid results from the fact that the shearing stress is greater on one side of the element than on the other. This in turn is due to a progressive change with distance of the velocity gradient of the fluid flow. This net force due to viscous resistance has the same direction as the fluid velocity.

We orient our element of volume with its sides parallel to the flow lines of the fluid, and its ends perpendicular to these flow lines in such a manner that the maximum rate of fluid shear occurs between one pair of its faces.

We choose axes such that the maximum shearing stress is parallel to the X-axis and acts upon the plane perpendicular to the Z-axis.

Let τ_{zx} be the shearing stress on one face, and let this increase to $\tau_{zx} + \frac{\partial \tau_{zx}}{\partial z} \cdot dz$ on the opposite face. The resultant force will be the

difference between the products of these two stresses by the areas of the faces acted upon, and will be directed toward the positive end of the X-axis.

$$f_r = \left(\tau_{zx} + \frac{\partial \tau_{zx}}{\partial z} \cdot dz\right) dxdy - \tau_{zx} dxdy$$
$$= \frac{\partial \tau_{zx}}{\partial z} \cdot dxdydz = \frac{\partial \tau_{zx}}{\partial z} \cdot dV, \tag{38}$$

where f_r is the net force due to viscous resistance acting upon the volume dV.

By equation (35) we can express the shearing stress in a fluid in terms of the viscosity and the gradient of the velocity, since

$$\tau_{zx} = \xi \frac{d\theta_{zx}}{dt},$$

$d\theta$ equals the ratio of the displacement in time dt of the upper layer over the lower, to the distance separating the two. Hence

$$d\theta_{zx} = \frac{dudt}{dz}, \text{ or } \frac{d\theta_{zx}}{dt} = \frac{du}{dz},$$

where u is the velocity in the X direction. Consequently

$$\tau_{zx} = \xi \frac{du}{dz}. \tag{39}$$

By combining (39) and (38) we obtain for f_r:

$$f_r = \frac{\partial \tau_{zx}}{\partial z} \cdot dV = \xi \frac{\partial^2 u}{\partial z^2} \cdot dV. \tag{40}$$

The force due to inertia is

$$f_i = dm \cdot a = \rho \cdot dV \cdot a, \tag{41}$$

and from (40) and (41) we obtain the ratio

$$\frac{f_i}{f_r} = \frac{\rho a \cdot dV}{\xi \cdot \frac{\partial^2 u}{\partial z^2} \cdot dV} = \frac{\rho a}{\xi \cdot \frac{\partial^2 u}{\partial z^2}}$$

If we resolve the members of the right hand term into their constituent elements and arrange them in a slightly different manner we get

$$\frac{f_i}{f_r} = \frac{\rho a}{\xi \cdot \frac{\partial^2 u}{\partial z^2}} = \frac{\rho \cdot \frac{v}{t}}{\xi \cdot \frac{v}{l^2}} = \frac{lv}{\frac{\xi}{\rho}} = \frac{lv}{\nu} = R, \tag{42}$$

where l is any characteristic length, v a characteristic velocity, ν the ratio of the viscosity to the density, known as the *kinematic viscosity*, and R, the numerical value so determined, is a constant known as Reynold's Number.

Reynold's Number, which derives its name from Osborne Reynolds,[20] who in his pioneer experiments first appreciated its significance, is one of the most important quantities in fluid mechanics. If we have a fluid model dynamically similar to the original we get for the model ratio

$$\frac{R_2}{R_1} = \frac{\lambda\eta}{\frac{\psi}{\delta}} = \frac{\lambda^2\tau^{-1}}{\lambda^2\tau^{-1}} = 1. \tag{43}$$

Consequently, for a geometrically similar model of a fluid, the canons of dynamic similarity are entirely satisfied when the Reynold's Number for the model is the same as that for the original, the Reynold's Number being, of course, based upon corresponding velocities and lengths in the two cases.

The Reynold's Number, in any given case, increases with the velocity of the fluid, and experiment has shown that in each case at a certain critical value of Reynold's Number there is a transition from laminar to turbulent motion. This transition occurs in the model employing a totally different fluid from the original at identically the same value of Reynold's Number.

This criterion enables us to perform model experiments using quite different fluids in the model from that in the original. A model experiment of an airplane, for example, can be made using water for the fluid and still obtain results that are quantitatively correct for an original in air.

The smaller the Reynold's Number in any particular instance, the greater the stability of the fluid motion.

Strength and Plasticity.—So far we have spoken of the properties of elasticity and of viscosity. Elastic deformation is characterized by the fact that when a body is subjected to a stress it undergoes a finite strain proportional to the stress applied, and that when the stress is released the strain disappears and the body regains its original form. A body is said to be viscous, on the other hand, if when a shearing stress is applied the body deforms continuously at a rate proportional to the applied shearing stress, approaching zero as the stress approaches zero.

It has already been stated that the most general state of stress at a

[20] Osborne Reynolds: *An experimental investigation of the circumstances whether the motion of water will be direct or sinuous, and of the law of resistance in parallel channels*, Royal Soc. London, Philos. Tr. (1883) *or* Sci. Pap., vol. 2, p. 51.

point inside a body is characterized by the three principal stresses, σ_1, σ_2, and σ_3, at that point. It can be shown that the maximum shearing stress occurs on a pair of conjugate planes parallel to the axis of intermediate stress, σ_2, and bisects the angles between the greatest and the least principal stresses, σ_1 and σ_3. The magnitude of the maximum shearing stress is equal to half the difference between the greatest and the least principal stresses.

$$\tau_{max} = \frac{\sigma_1 - \sigma_3}{2}.$$

When

$$\sigma_1 = \sigma_2 = \sigma_3,$$
$$\tau_{max} = 0,$$

and all shearing stresses vanish.

When an elastic solid is subjected to unequal principal stresses it deforms elastically, but as the difference between the greatest and the least principal stresses is made continuously larger a state of shear strain is reached beyond which the body will not return to its original form upon release of the stress. The body is then said to have reached its *elastic limit*. If the stress continues to be further increased the body will fail by slip, by rupture, by brittle fracture, or else a *yield point* will be reached and the body will flow *plastically*.

The particular stress combination under which a given material fails by fracture or flows plastically is said to be its *strength*. Strength is not a constant but is a dependent variable; it is a function of the three principal stresses and of the temperature. The strength of a given material decreases as the temperature is increased, becoming zero in the vicinity of the melting point. As the intermediate principal stress, σ_2, is increased the strength of a material as measured by the difference between the greatest and the least principal stresses, σ_1 and σ_3, at which fracture or plastic deformation occurs, increases.

The most common tests made in testing laboratories to determine the strengths of materials are the application of simple tension, longitudinal compression, and shear stress. These give the tensile strength, the crushing strength, and the shear strength, respectively. By a variety of more elaborate experiments Adams and associates, von Kármán, Bridgman, Griggs, and others have measured the strengths of rocks under high confining pressures. The strengths of rocks have been increased by a factor of 10 times or more in this manner.

It is thus clear that the strength of a material is neither a simple nor a well-defined property. For our purposes it is sufficient, however, that strength is measured in terms of stress. If in the original, failure occurs

in a given manner at a certain stress, then in the model, for similarity, failure must occur in the same manner at the corresponding value of the stress. Hence the model ratio of strength must be the same as that of stress.

$$\frac{(\text{strength})_2}{(\text{strength})_1} = \frac{(\text{stress})_2}{(\text{stress})_1} = \sigma = \mu\lambda^{-1}\tau^{-2}. \tag{44}$$

Many substances, the malleable metals for example, deform plastically at ordinary temperatures and zero intermediate stress. Most rocks, however, fail under similar circumstances by brittle fracture. The above-mentioned experiments with rocks under high confining stresses have all shown that rocks, brittle under ordinary conditions, become plastic under high enough confining stress.

The most important experiments of this kind are those recently reported by Bridgman.[21] A bar of hardened steel called the "anvil" was placed between two opposed hardened steel pistons. Powdered specimens of materials to be tested were placed between the pistons and the anvil. The pressure on the pistons was increased and the anvil was periodically rotated.

For low pressures of the pistons the specimen would slip at the surface of contact and behave as a rigid body. As the pressure on the pistons was increased, however, a point would be reached at which the specimen would shear internally as the anvil was rotated. By knowing the area of the specimen and measuring the torque necessary to cause shear, the shearing strength of the material could be computed. Bridgman found that the shearing strength of the materials tested increased as the pressure on the pistons was increased. The maximum pressure obtained was 50,000 kg./cm.^{2}, which is equivalent to a depth of 166 km. beneath the earth's surface. At this highest pressure the shearing strength of most materials was increased by a factor of 10 times or more over its value under ordinary conditions.

The manner of failure or of flow is also important. Many substances, once they reached the plastic state, would flow continuously and could be made to shear an indefinite amount by continued rotation of the anvil. Other substances could never be made to flow continuously. In the latter, shearing stress would build up to a maximum and then be released by sudden jumping and crackling as the anvil was rotated.

On a few of the substances treated the shear stress increased about 10 times as the velocity of shear was increased 10,000 times. In most of the materials, however, there was no measurable change of the shearing strength as a function of the rate of shear.

[21] P. W. Bridgman: *Shearing phenomena at high pressure of possible importance to geology*, Jour. Geol., vol. 44 (1936) p. 653-669.

The fact that in plastic deformation the shear stress necessary to produce plastic shear is very nearly independent of the rate of shear differentiates this type of deformation from viscous shear, where the stress is proportional to the rate of shear. Plastic bodies undergo elastic deformation until, under the conditions obtaining, the shear stress equals the shear strength; then they shear plastically at constant values of the shear stress, which is essentially independent of whether the shear movement is slow or rapid.

This independence between the shear stress and the rate of shear in plastic deformation makes it possible to increase the speed of a plastic process without materially altering the magnitudes of the stresses involved. This enables us to distort the time factor in models of mountain making so that in the model an amount of deformation may be made in a few hours which otherwise should have required very much longer time, without any effective loss of similarity—a point that we shall return to later.

Elastico-viscosity.—Besides solid, fluid, and plastic bodies there are materials such as sealing wax which deform elastically to stresses of short duration and yet deform viscously to extremely minute shearing stresses of long duration. Maxwell [22] has called these *elastico-viscous* materials. Under large enough stresses these bodies fail like ordinary solids. Sealing wax fractures brittlely.

In such materials the coefficient of viscosity is quite high. The rate of deformation cannot be speeded up beyond a certain amount because when the stress reaches a certain critical value the substance fails by fracture. By analogy with other substances it is highly probable that under suitable stress an elastico-viscous material can be made to fail by plastic flowage.

There is much evidence to indicate that rocks are elastico-viscous materials of very high viscosity. The transition to the plastic state occurs under suitable stress differences and we then get more rapid deformations such as mountain making. Viscous movements are exemplified in the post-Glacial uplift of the Great Lakes region and of Scandinavia.

RELATION BETWEEN MODEL RATIOS AND PHYSICAL DIMENSIONS

Fundamental Units.—In mechanics, all quantities are measured by appropriate combinations of three fundamental units: *mass, length,* and *time.* Quantities other than these fundamental ones are said to be *derived quantities.* The *physical dimensions* [23] of a quantity are expressed by the number of times and manner in which the fundamental quanti-

[22] J. Clerk Maxwell: *On the dynamical theory of gases,* Philos. Mag., vol. 35 (1868) p. 134.
[23] P. W. Bridgman: *Dimensional analysis,* revised ed. (1931) Yale Univ. Press.

ties are involved in the measurement of that quantity. A *dimensional expression* is an expression, usually housed in brackets, of the symbols M, L, and T, representing mass, length, and time, respectively, where each symbol has an exponent signifying the number of times it occurs. A positive exponent signifies that the symbol belongs in the numerator; a negative exponent signifies the denominator; and a fraction signifies a root. For example

$$\begin{array}{ll} [\text{Dimension of quantity}] & = [M^a L^b T^c]; \\ [\text{Velocity}] & = [M^0 L T^{-1}] = [L T^{-1}]; \\ [\text{Force}] & = [M L T^{-2}]. \end{array} \tag{45}$$

A physical equation is valid only provided every term has the same dimensions as every other term. It is physically not permissible, for example, ever to do the equivalent of adding horses to apples—a procedure employed regularly and as a matter of course in economic computations.

It will be noted, and can be shown to be of general validity, that when an expression for the model ratio, in terms of μ, λ, and τ is of the form $\mu^a \lambda^b \tau^c$, the dimensional expression in terms of M, L, and T is of the form $M^a L^b T^c$. This important fact enables us to write by inspection the model ratios for all mechanical quantities from their dimensional formulas. This is done in Table 2 for the more commonly used quantities.

Alternative Fundamental Units.—Frequently it is convenient to express the model ratio in terms of ratios of density, velocity, viscosity, acceleration, or strength, taken as fundamental rather than in terms of those of length, mass, and time. Any three such ratios will serve equally well provided only that they contain the ratios of mass, length, and time.

Transition from one set of ratios, taken as fundamental, to any other may be made quite readily by setting up the appropriate equations of transformation. For example, suppose we wish to take the ratios of length, density, and acceleration as fundamental. For equations of transformation, we have

$$\left.\begin{array}{ll} \lambda = \lambda & \lambda = \lambda \\ \delta = \mu\lambda^{-3} \quad \text{or} & \mu = \delta\lambda^3 \\ \gamma = \lambda\tau^{-2} & \tau = \lambda^{\frac{1}{2}}\gamma^{-\frac{1}{2}} \end{array}\right\}. \tag{46}$$

By means of these equations any quantity may be transformed from either set of units to the other.

SPECIAL CASES

General Statement.—We have now covered the essential theory necessary to enable us to compute the properties of any kind of a mechanical model—rigid, plastic, or fluid—provided the properties of the original are known. Before we proceed to the computation of actual mod-

els, however, there are a number of aspects of the general theory to which attention paid now will forestall serious difficulties later on.

In setting up our basic criterion of dynamic similarity—that all forces of like kinds must be proportional—we chose the ratio of the forces due

TABLE 2.—*Model ratios of mechanical quantities*

Quantity	Dimensional Formula	Model Ratio
Angle	L^0	1
Area	L^2	λ^2
Volume	L^3	λ^3
Curvature	L^{-1}	λ^{-1}
Frequency	T^{-1}	τ^{-1}
Velocity	LT^{-1}	$\lambda\tau^{-1}$
Acceleration	LT^{-2}	$\lambda\tau^{-2}$
Angular velocity	T^{-1}	τ^{-1}
Angular acceleration	T^{-2}	τ^{-2}
Density	ML^{-3}	$\mu\lambda^{-3}$
Momentum	MLT^{-1}	$\mu\lambda\tau^{-1}$
Moment of momentum	ML^2T^{-1}	$\mu\lambda^2\tau^{-1}$
Angular momentum	ML^2T^{-1}	$\mu\lambda^2\tau^{-1}$
Force	MLT^{-2}	$\mu\lambda\tau^{-2}$
Torque	ML^2T^{-2}	$\mu\lambda^2\tau^{-2}$
Work and energy	ML^2T^{-2}	$\mu\lambda^2\tau^{-2}$
Power	ML^2T^{-3}	$\mu\lambda^2\tau^{-3}$
Action	ML^2T^{-1}	$\mu\lambda^2\tau^{-1}$
Stress	$ML^{-1}T^{-2}$	$\mu\lambda^{-1}\tau^{-2}$
Strain	L^0	1
Elastic modulus	$ML^{-1}T^{-2}$	$\mu\lambda^{-1}\tau^{-2}$
Viscosity	$ML^{-1}T^{-1}$	$\mu\lambda^{-1}\tau^{-1}$
Kinematic viscosity	L^2T^{-1}	$\lambda^2\tau^{-1}$
Gravitational constant	$M^{-1}L^3T^{-2}$	$\mu^{-1}\lambda^3\tau^{-2}$

to inertia, $\mu\lambda\tau^{-2}$, as the standard to which all other forces have been made to conform. We encountered at once, however, a certain duality in the controlling factor of the force ratio owing to the fact that where both the model and the original are on the earth's surface, the same gravity, g, acts in both cases. This at once forces us to set the ratio of acceleration, $\lambda\tau^{-2}$, equal to unity, whereupon λ becomes equal to τ^2 for all such experiments. Consequently under such restrictions λ and τ are no longer independent ratios, but rather are so related to each other that the choice of a value for either uniquely determines that for the other.

Suppose, for example, that we wish to make a model of some earth phenomenon which required a million years. We wish to perform our

experiment within a few hours. The time ratio would be of the order of 10^{-9}.

In such an instance, λ, which is equal to τ^2, becomes 10^{-18}, and if the original were the earth itself, the model would have to be about the size of a molecule.

Suppose that on the other hand the original is of the order of 100 kilometers in length and we wish the model to be a convenient size for laboratory use. A value for λ of 10^{-5} would give a length of 100 centimeters for the model. Then τ, which is $\lambda^{\frac{1}{2}}$, would become $10^{-\frac{5}{2}}$. If the original time were 10^6 years, the corresponding time for the model would be $10^{3.5}$ or 3,160 years.

A somewhat similar difficulty is encountered in a problem such as this: It is known that a small flywheel made of a certain material will explode at a certain angular velocity. It is desired to know the corresponding velocity of a geometrically similar, larger flywheel made of the same material.

Our strict theory of similarity tells us nothing whatever about such a problem. It tells us instead that if we change the size of the wheel we also must change its strength, and hence its materials, or else the two cases will not be dynamically similar.

It is this objectionable inflexibility of the strict theory that we now must find a way to circumvent, for otherwise in models of geological phenomena either length or time will always be getting out of bounds. If we choose the time ratio to suit our experiment the length ratio becomes such as to render the model sub-microscopic in size; or if we choose the length ratio for convenience, the time ratio becomes such that the time required is much longer than that available for making the experiment.

A key to the solution of the difficulty is found when we return to first principles and scrutinize carefully the requirement for dynamic similarity—that all forces of a like kind must be proportional. We have already remarked that we have a duality of control in that there are two body forces, one due to inertia and the other due to gravity, both of which must simultaneously be satisfied. In general, both these forces do exist and, if they exist, both must be satisfied. The ratio of the forces due to inertia is $\mu\lambda\tau^{-2}$ and that due to gravity is $\mu\frac{g_2}{g_1}$ which on the earth's surface is equal to μ. Consequently, in such cases

$$\mu\lambda\tau^{-2} = \mu, \quad \text{or} \quad \lambda\tau^{-2} = 1,$$

as remarked before.

Inertial Forces Negligible.—Suppose, however, that in certain instances both forces do not exist simultaneously or, if they do, that one of them

is negligible in magnitude as compared with the other. Consider for example a static structure in which the acceleration and hence the force due to inertia is zero; or a slowly moving viscous or plastic body in which the acceleration is so nearly zero as to produce forces of only infinitesimal magnitude. In such cases, while there is no incorrectness involved in basing the ratio ϕ of the forces upon the inertial ratio deter-

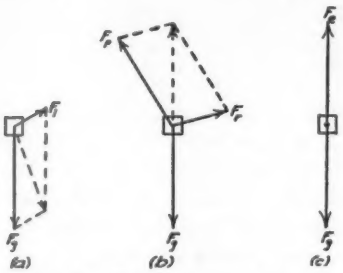

FIGURE 15.—*Special cases of force equilibrium*

(*a*) As force due to inertia approaches zero, the resultant body force approaches that due to gravity; force equilibrium without inertia in (*b*) fluid and (*c*) elastic body.

mined by $\mu\lambda\tau^{-2}$, dynamic similarity no longer depends upon our doing so. In this case the equilibrium of forces for elastic bodies becomes

$$F_g + F_e = 0, \tag{47}$$

and

$$F_g + F_r + F_p = 0, \tag{48}$$

for viscous and plastic bodies.

Obviously in such cases the real controlling factor is not the body force due to inertia, for this is practically non-existent, but is instead that force due to gravity which on the earth's surface is quite independent of the model ratio of time employed in the particular experiment. Consequently, in cases where the acceleration is so small that forces due to inertia are negligible the canons of dynamic similarity are entirely satisfied if we choose the model ratios μ, λ, and τ arbitrarily and make all forces conform to the ratio

$$\phi = \mu\gamma_g, \tag{49}$$

where γ_g is the model ratio of the acceleration due to gravity and may, for the earth's surface, be set with negligible error equal to unity. Then

$$\phi = \mu = \delta\lambda^3. \tag{50}$$

The same result is achieved if we factor $\lambda\tau^{-2}$ out of all the general model ratios for the derived quantities, such as stress, strength, viscosity, and the like when and where needed.

For example:

Quantity	Strict model ratio	Model ratio neglecting inertia
Force	$\mu\lambda\tau^{-2}$	μ
Stress	$\mu\lambda^{-1}\tau^{-2}$	$\mu\lambda^{-1}$
Strength	$\mu\lambda^{-1}\tau^{-2}$	$\mu\lambda^{-1}$
Viscosity	$\mu\lambda^{-1}\tau^{-1}$	$\mu\lambda^{-1}\tau$

Since Reynold's Number was derived from the ratio of the inertial to the resistive forces it will be noted that as the inertial forces become negligible, Reynold's Number approaches zero. We have already seen that for the general case, dynamic similarity requires that the Reynold's Number for the model be equal to that for the original. This placed us under the severe restriction that

$$\frac{\frac{\lambda\eta}{\psi}}{\delta} = 1,$$

as a necessary requirement for similarity. In those cases where the inertial forces are negligibly small, we can now violate this restriction and obtain dynamic similarity when the Reynold's Numbers are no longer equal in the two cases, though quite small.

Gravitative Forces Negligible.—Another special case is the one exemplified by the flywheel in which the acceleration due to gravity is negligible compared to that of the motion of the body. In such an instance we may completely neglect the forces due to gravity and base the force ratio entirely upon the ratio of the forces due to inertia. This gives us throughout equations identical to those we have already derived from the strict theory with the exceptions that the relation

$$\lambda\tau^{-2} = 1,$$

which was formerly imposed by gravity, is now no longer required and can, when gravity is negligible, be completely ignored, λ and τ being taken as independent variables.

Resistive Forces Negligible.—We have still another special case when the forces due to resistance are so small in comparison with those due

to gravity, to inertia, and to pressure, that their effect is negligible. Such cases are to be found in the flow of fluids of small viscosity through large orifices, in the flow of water over a weir, and in the movement of boats at high velocities.

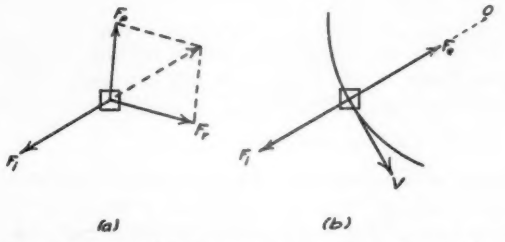

FIGURE 16.—*Inertial force large compared with gravitational force*
(a) fluid, (b) rotating solid.

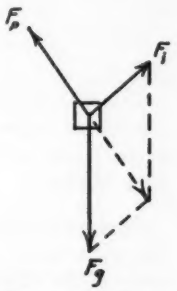

FIGURE 17.—*Force due to resistance negligible*

Whereas, for the general case, dynamic similarity could be achieved only when the proper viscosity ratio

$$\psi = \mu\lambda^{-1}\tau^{-1},$$

was satisfied, in this case the ratio of the viscosities may be ignored, or by using the same fluid in both cases, be made equal to unity.

Similarity Based on Resultant Body Forces.—An effective way of changing the force ratio of static structures is to base the model ratio of force on the vector sum of the forces due to accelerated motion and those due to gravity, without regard to the ratio between these two kinds of body force. This enables one to obtain similarity for the case in which the original is a static structure acted upon by gravity, and the model is a geometrically similar static structure being whirled in a centrifuge.

The model ratio of force then is

$$\phi = \frac{(\text{body forces})_2}{(\text{body forces})_1} = \frac{m_2(\vec{g} + \vec{a})}{m_1 g} \cong \mu \cdot \frac{a}{g}.$$

The length ratio λ is arbitrary. The time ratio τ is not involved. The stress ratio is given by

$$\sigma = \phi\lambda^{-2} = \mu\frac{a}{g} \cdot \lambda^{-2} = \delta\lambda \cdot \frac{a}{g}.$$

In particular if the stress in the model is made equal to that in the original, the same materials can be used in both. Then

$$\sigma = 1,$$
$$\delta = 1,$$

and

$$a = \frac{g}{\lambda}.$$

This tells us that if the model is reduced by a length ratio of 1/1000, a must be made equal to $1000g$, if the same materials are used. This is the basis of Bucky's [24] experiments employing a centrifuge to test mine models.

Summary of Special Cases.—To summarize, we find that the mechanical properties of any model can be completely specified, when the properties of the original are known, in terms of the three fundamental model ratios μ, λ, and τ of mass, length, and time. We distinguish the general case and three special cases as follows:

1. General case.

   Independent fundamental ratios: μ, λ, and τ.

$$\phi = \frac{f_{i2}}{f_{i1}} = \frac{f_{g2}}{f_{g1}} = \mu\gamma = \mu\gamma_g.$$

Therefore,

$$\gamma_g = \gamma = \lambda\tau^{-2};$$
$$\phi = \mu\lambda\tau^{-2}.$$

2. Special cases.

   A. *Both the model and the original are at the earth's surface.*

   Fundamental model ratios: μ, λ, and τ.

$$\frac{g_2}{g_1} = \gamma_g = 1 = \lambda\tau^{-2}$$

Therefore,

$$\tau = \lambda^{\frac{1}{2}}.$$

Hence, the independent model ratios reduce to two: μ and λ, or μ and τ.

   B. *The forces due to inertia are negligible compared with those due to gravity; Reynold's Number is small.*

   Fundamental model ratios: μ, λ, and τ.

$$\phi = \mu\gamma_g;$$
$$\lambda\tau^{-2} \neq \gamma_g;$$
$$\gamma_g = 1 \text{ at earth's surface;}$$
$$\lambda \text{ independent of } \tau.$$

[24] P. B. Bucky: *The use of models for the study of mining problems*, Am. Inst. Min. Metall. Eng., Tech. Pub. 425 (1931).

C. *Force due to gravity is negligible compared with that due to inertia, or else is eliminated from consideration by the arrangement of the experiment.*

Fundamental model ratios: μ, λ, and τ.

$$\lambda \text{ is independent of } \tau;$$
$$\phi = \mu\lambda\tau^{-2}.$$

D. *Force due to viscous resistance negligible as compared with other forces.*

Fundamental model ratios: μ, λ, and τ.

$$\tau = \lambda^{\frac{1}{2}} \text{ at earth's surface;}$$
$$\psi \text{ independent of } \mu, \lambda, \text{ and } \tau; \text{ may be equal to unity.}$$

E. *Similarity based upon model ratio of total body forces.*

Fundamental model ratios, μ, λ, and ϕ.

$$\phi = \frac{a_2}{g_1}$$

when

$$a_1 = 0,$$

$$g_2 = g_1,$$

and

$$a_2 >> g_1.$$

ILLUSTRATIVE EXAMPLES

Introduction.—Before proceeding with geologic problems it perhaps would be useful to demonstrate the theory we have developed by applying it to a few simple illustrative examples.

Viscous Fluid.—Suppose we have a barrel of tar and a cannon ball. The cannon ball is placed upon the tar and sinks slowly to the bottom. Using this for our original, we wish to construct a dynamically similar model on a reduced scale. What must the properties of the model be?

We proceed first by means of the general theory. μ, λ, and τ are our fundamental model ratios. We derive the remaining ratios that are of interest to us:

Viscosity ratio,	$\psi = \mu\lambda^{-1}\tau^{-1}$;
Velocity ratio,	$\eta = \lambda\tau^{-1}$;
Density ratio,	$\delta = \mu\lambda^{-3}$.

We now make use of two special circumstances: that gravity is the same in the model and in the original, and that the motion is so slow

that the forces of inertia may be neglected. This corresponds to Special Case B already discussed. μ, λ, and τ are independent, and ϕ is equal to μ. So long as the inertial forces may be neglected in the model, dynamic similarity is achieved when

$$\psi = \mu\lambda^{-2}\tau,$$
$$\eta = \lambda\tau^{-1},$$
$$\delta = \mu\lambda^{-3}.$$

If the density of the model is made equal to that of the original then δ becomes equal to unity, and

$$\mu = \lambda^3,$$

whence

$$\psi = \lambda\tau,$$
$$\eta = \lambda\tau^{-1}.$$

Thus for a given length scale λ we can vary the ratio of viscosities ψ widely and still obtain dynamic similarity. This enables us to adjust the time scale to our convenience.

Flywheel.—A small flywheel explodes at a certain critical angular velocity. At what velocity would a dynamically similar large flywheel explode?

If we solve this for the case where gravity is the same in the model and the original we choose model ratios μ and λ and determine τ by the fact that

$$\lambda\tau^{-2} = 1.$$

For the derived model ratios we obtain for:

Strength:	$\sigma = \mu\lambda^{-1}\tau^{-2} = \mu\lambda^{-2}$;
Angular velocity:	$\frac{\omega_2}{\omega_1} = \tau^{-1} = \lambda^{-\frac{1}{2}}$.

That is, the angular velocity at which the wheel will explode changes with the size of the wheel as $1/\sqrt{\lambda}$, but by the strict theory this is true only when the strength of the material has been changed by a factor $\mu\lambda^{-2}$.

In this case, however, the forces due to gravity are negligible and so by neglecting gravity, the ratios λ and τ again become independent.

The model ratios then become

Strength:	$\sigma = \mu\lambda^{-1}\tau^{-2}$;
Angular velocity:	$\frac{\omega_2}{\omega_1} = \tau^{-1}$.

Consequently, we may use the same materials in both wheels, giving for the strength and density ratios:

$$\sigma = \mu\lambda^{-1}\tau^{-2} = 1;$$
$$\delta = \mu\lambda^{-3} = 1;$$

from which

$$\lambda = \tau,$$

and the ratio of angular velocities becomes

$$\frac{\omega_2}{\omega_1} = \tau^{-1} = \frac{1}{\lambda}.$$

Static Structure.—For a static structure all the accelerations due to motion are zero. The model is completely determined by the model ratios μ and λ,

$$\phi = \mu,$$

and

$$\sigma = \mu\lambda^{-2},$$

determine the ratio of the forces, and of the stresses and the strengths of the materials.

If the ratio δ for density is employed instead of μ,

$$\delta = \mu\lambda^{-3},$$
$$\phi = \delta\lambda^{3},$$
$$\sigma = \delta\lambda.$$

When δ is unity, corresponding to the fact that the density is the same in both the model and the original,

$$\sigma = \lambda.$$

This tells us that if a structure is increased, for example, 100 times in linear dimensions the strength of its materials must also be increased 100 times if it is to support a corresponding load. Since in actual structures the strengths of the materials rarely are increased by such an amount, the resulting weakness must be compensated by departure from geometrical similarity. For this reason, small animals have spindley legs and large animals have massive legs.

Flow of Water Over Weir.—The constants of a model of water flowing over a weir are to be determined.

Solving for the general case first, we choose μ, λ, and τ, the latter two being related by $\lambda\tau^{-2} = 1$. The derived model ratios that are of interest then become:

Velocity:	$\eta = \lambda\tau^{-1} = \lambda^{\frac{1}{2}}$,
Density:	$\delta = \mu\lambda^{-3}$,
Viscosity:	$\psi = \mu\lambda^{-1}\tau^{-1} = \mu\lambda^{-\frac{1}{2}}$,
Discharge:	$\frac{D_2}{D_1} = \lambda^{3}\tau^{-1} = \lambda^{\frac{5}{2}}$.

Since in this case it is impossible to have both δ and ψ set equal to unity simultaneously, this tells us that we can have the model dynamically similar to the original only provided a liquid of different kinematic viscosity from that of the original is used in the model.

We may avoid this restriction by noting that the frictional forces are negligible compared with the remaining forces (Special Case D). Then we can ignore the viscosity altogether and use water in the model as well as in the original. We obtain in this manner:

Velocity:	$\eta = \lambda\tau^{-1} = \lambda^{\frac{1}{2}}$,
Density:	$\delta = \mu\lambda^{-3} = 1$,
Discharge:	$\frac{D_2}{D_1} = \lambda^{3}\tau^{-1} = \lambda^{\frac{5}{2}}$,
Viscosity:	$\psi = 1$.

This tells us that for fluids of low viscosity flowing over geometrically similar weirs (or through orifices), the discharge increases as the length ratio raised to the power 5/2, where the *same* fluid is used in both instances.

MODELS OF GEOLOGIC STRUCTURES

PROCEDURE

General Statement.—In the light of the foregoing development we are now prepared to answer a question propounded earlier in this paper: What should the mechanical properties of the materials of a geological model be in order that the model may give results which are dynamically similar to the original? We shall derive the answer to this question for a number of cases of geological importance. By applying a similar procedure anyone can solve any other particular problem that may be of interest to himself.

Before proceeding with this, however, perhaps it should be made clear that in this paper it is only intended to show how to determine the necessary specifications for the geological models in order that they may behave in a manner similar to the original. In many cases it may be difficult or impossible to obtain materials having the properties specified. Fortunately the theory enables us to vary the specifications somewhat widely, so that with a variety of materials it will frequently

be possible to adjust the model ratios until an approximate fit is obtained. The present paper is only a guide to the experimenter to give him a means of knowing to what extent his model materials do satisfy the requirements of similarity. It does not purport to foresee or to solve the experimental difficulties to be encountered in actually building a model.

There is still another thing about geologic models that perhaps should be mentioned. That pertains to boundary conditions, and the manner in which the deformation is to be induced. The geologic model is likely to have boundaries that do not exist in the original. This may not be serious if these boundaries are remote from the region of the model being studied. The method of inducing deformation in the model necessarily involves assumptions regarding the stress relations in the earth, and the model accordingly will vary with what the experimenter assumes to be the case. Variation of these assumptions should then lead to variations of the experiment. These, however, are problems that the individual experimenter must solve for himself.

Granite Cube.—For purely illustrative purposes let us begin with the simplest possible kind of problem. Let us imagine a cube of flawless granite, 20 kilometers to the side, resting upon a plane, horizontal, rigid base. We wish to determine the properties of a dynamically similar model 20 centimeters to the side.

This is a static problem and hence a special case of the more general theory. The model ratios of interest to us are:

Fundamental ratios:	μ and λ
Acceleration:	$\gamma = 0;\ \gamma_g = 1$
Density:	$\delta = \mu\lambda^{-3}$
Force:	$\phi = \mu\gamma_g = \mu$
Stress and strength:	$\sigma = \phi\lambda^{-2} = \delta\lambda$

The ratios of length, density, and strength are the ones that concern us most, for they determine the size, the density, and the strength of the model materials. The length ratio λ is determined from the problem as given to be 10^{-5}. The density of granite is 2.7; that of convenient model materials would range between about 1.5 and 2.7 which would determine the value of δ to be in the range from 0.5 to 1.0. From these the model ratio of strength may be determined. If δ is 1.0, σ will be equal to λ, or 10^{-5}. If δ is 0.5, σ will be 5×10^{-6}.

For simplicity let us make the density of the model the same as that of granite. Then in this case δ will be unity and σ will be equal to λ, or to 10^{-5}. This means that the strength of the model materials will have to be 10^{-5} of the strength of granite. Taking the crushing strength of granite to be 2×10^9 dynes/cm^2 or 29,000 lbs./in^2 we find that the

model material must be a brittle substance that will crush under a stress of 2×10^4 dynes/cm^2 or 0.29 lbs./in^2.

Since the pressure at the base of the model would be 5.3×10^4 dynes/cm^2, which is about 2.7 times as great as the crushing strength of the material, it follows that the model would be too weak to stand unsupported and would collapse under its own weight.

That this is not an absurd result may readily be seen when one computes the pressure at the base of the original cube of granite, which will also be found to be 2.7 times as great as its crushing strength so that the original would likewise collapse under its own weight. Therefore the strength we have chosen is the proper one if the model is to behave in a manner dynamically similar to the original.

Mountain Range.—We take up now a larger unit. We choose a mountain range for the original and again we wish to know the model ratios and essential model constants. This is the same kind of a special case as that discussed for the granite cube. Aside from occasional earthquake motions of small magnitude, the accelerations are sensibly zero and may be neglected. The model ratios are the same (algebraically) as in the case of the cube, and in particular the ratio for strength is

$$\sigma = \delta\lambda.$$

For the original, suppose the width is 200 kilometers, the density 3.0, and the strength of the materials 2×10^9 dynes/cm^2, or the strength of granite under surface conditions. For the model let the width be 1 meter and the density 1.5. This gives for λ a value of 5×10^{-6}, and for δ a value of 0.5. Then σ, which is $\delta\lambda$, becomes 2.5×10^{-6}. Consequently the strength of the model would have to be:

$$\text{Model strength} = \sigma \times (\text{Strength of original}) = 5 \times 10^3 \text{ dynes/cm}^2.$$

Since on such a scale rocks are known to deform principally by plastic flowage, the model material would need to be more plastic than brittle.

The above figure for strength is taken merely as an approximation and for illustrative purposes. Actually the strength of rocks is known to increase with increase of confining stress—that is, with depth in the earth. The strength also, however, decreases with increase of temperature, which also increases with depth. Consequently the actual strength of rocks at a given depth within the earth is unknown except by rather indirect considerations.

When one is actually making a model each part of it should have the strength so chosen as to give the model the same strength inhomogeneities as are known or assumed to exist in the original. Neglecting for present purposes the inhomogeneities it is significant that the value of

5.0×10^3 C. G. S. for the over-all strength of the model materials corresponding to the case given is the right order of magnitude. This corresponds to a material so weak that a cube of it (density, 1.5) larger than 3.3 centimeters to the side could not support its own weight. Butter at ice box temperature would be much too strong. Vaseline or very soft clay would be more nearly correct.

Post-Glacial Uplift.—A somewhat different kind of problem is that exemplified by the gradual uplift that has been occurring in the regions of the Great Lakes and of the Scandinavian Peninsula since the retreat of the last ice sheet. Nansen [25] and Daly [26] have presented data which appear to indicate that both in North America and in Europe this uplift has occurred in a manner similar to the uplift of the surface of a barrel of tar following the removal of a weight. If this be so then the movement would occur as true viscous flow in an elastico-viscous medium having very high viscosity.

For our purposes we shall make the assumption that the departure from sphericity in the areas affected by post-Glacial uplift is slight enough that they may be represented by a model having a plane surface. We shall also assume that the earth has uniform viscosity to great depth.

Our model will then consist of a large vessel of viscous material, the surface of which has been depressed and is later returning to equilibrium. If we knew the over-all viscosity of the earth we could, from the model ratio of viscosity, determine the proper viscosity for the model. If, however, we regard the earth viscosity as an unknown to be determined, the model theory should tell us how to compute the earth viscosity from that of our model. The results in either case will not be exact because of the simplifying assumptions made, but they should give us the correct order of magnitude.

We choose in this case the fundamental model ratios μ, λ, and τ of length, mass, and time. We note that we are dealing with a fluid type of motion having an extremely small Reynold's Number and that all accelerations are sensibly zero. Consequently the inertial forces are negligible as compared with the resistive forces. Under these circumstances μ, λ, and τ are to be regarded as mutually independent model ratios from which we write the desired derived ratios:

Acceleration:	$\gamma = 0;\ \gamma_g = 1,$
Force:	$\phi = \mu\gamma_g = \mu,$
Velocity:	$\eta = \lambda\tau^{-1},$
Stress:	$\sigma = \mu\lambda^{-2},$
Density:	$\delta = \mu\lambda^{-3},$
Viscosity:	$\psi = \sigma\tau = \mu\lambda^{-2}\tau = \delta\lambda\tau.$

[25] Fridtjof Nansen: *The earth's crust, its surface, forms, and isostatic adjustment* (1928) Oslo.
[26] R. A. Daly: *Our mobile earth* (1926) New York.

The ratios concerning us most are those of density, length, time, and viscosity, and we see that that of viscosity is determined when the other three are chosen.

The radii of the depressed regions both in North America and in Europe are of the order of 1,000 kilometers. In the model, to avoid boundary effects, it is necessary that the viscous medium be of much greater extent than just the depressed area. In order to do this and not have the model inconveniently large, suppose we give the depressed area a radius of 10 centimeters. This would give a value for λ of 10^{-7}.

The total time elapsed since the beginning of the ice retreat is unknown but is about 25,000 to 35,000 years. What is more important for the present purposes is, however, the time interval between successive amounts of uplift as shown by old shore lines. This appears to be known with some degree of accuracy in certain cases. The ice retreat itself required some 10,000 to 15,000 years.

To keep the model time within limits we may set τ equal to 10^{-7}. For this time ratio an amount of uplift requiring 10,000 years in the original would require slightly less than 9 hours in the model.

The density of the original may be taken as 3.0; that of the model materials will probably be near 1.0, giving a value of 0.33 for δ. From these ratios that of viscosity is obtained:

$$\psi = \delta\lambda\tau = 3.3 \times 10^{-15}$$

An experimental difficulty will be recognized at once when the amount of uplift is considered in comparison with horizontal distances involved. The total uplift is of the order of 250 meters as compared with a radial distance of about 1,000 kilometers. For a model with a 10-centimeter radius the depression, to scale, would be only 2.5×10^{-3} centimeters. This is a difficulty that would remain for any size of a model not too large to work with. Direct measurements of such small displacements would require probably some optical device such as an interferometer.

To avoid the difficulty inherent in small displacements, let us consider the possibility of introducing distortion by depressing the model by an amount equal to the correct depression multiplied by a factor β. So long as the amount of depression remains a small fraction of the radial distance the flow lines in the viscous material remain essentially unchanged. The velocity of flow, however, and also the rate of shear are directly proportional to the shearing stress and inversely to the viscosity. Since the viscosity in the model with distortion is unchanged, the shearing stress varies with the distortion.

Shearing stress in this case results from the fact that the model is not in static equilibrium. It is proportional to the displacement from the

equilibrium position. Consequently, if the amount the surface is depressed is increased by a factor β, with the radius of the depression kept constant, the shearing stress throughout the model will be increased by the factor β and the velocity of displacement will likewise be increased by the factor β.

Consequently, if the uplift in the undistorted model is a given amount in a certain time, the uplift in the model having distortion will be β times that amount. In particular, if in the undistorted model an uplift equal to any given fraction of the initial amount of the depression occurs in time t, an amount of uplift equal to the same fraction of its original depression will occur in the same time t in the model having distortion. In other words, the depressed surface approaches its equilibrium position as a negative exponential function of time, or

$$h = h_o e^{-at},$$

where h_o is the initial amount the surface was depressed, h the amount of depression still remaining at time t, e the base of natural logarithms, and a a constant.

When the distortion factor β is known, all that is necessary in order to obtain the correct displacement is to divide the observed displacement by β.

Once the model ratios are determined we may proceed in either of two directions, depending upon what is considered as known and what is unknown. Suppose that for the earth something is known of the amount of uplift and the time required, and it is desired to know the earth's coefficient of viscosity. Then, since

$$\psi = \frac{\xi_2}{\xi_1} = \delta\lambda\tau,$$

the viscosity of the earth is obtained by

$$\xi_1 = \frac{\xi_2}{\psi} = \frac{\xi_2}{\delta\lambda\tau}.$$

In this case δ and λ are chosen and τ and ξ_2, the model ratio of time, and the viscosity of the model, are determined by direct measurement.

If, on the other hand, τ and ξ_1 are regarded as known, ξ_2, the viscosity of the model, may be computed.

For the case given, we have already taken the model ratio of viscosity to be 3.3×10^{-15}. By semi-theoretical, semi-experimental considerations Haskell [27] concludes that the earth has a kinematic viscosity of 2.9×10^{21} C. G. S. Gutenberg cites figures of the same order of magnitude, or

[27] N. A. Haskell: *The motion of a viscous fluid under a surface load*, Physics, vol. 6 (1935) p. 265.

slightly less. This would give for the viscosity itself about 10^{22} C. G. S. Taking this figure, we would obtain for the model

$$\xi_2 = \xi_1\psi = 3.3 \times 10^7 \text{ C.G.S.}$$

This is about the viscosity of asphalt at 40 degrees Centigrade.

A lower viscosity for the model can be obtained by reducing the time required by the model, and hence further reducing the model ratio.

Salt Dome.—Although rock salt is crystalline and not fluid, there is much evidence that it deforms under long-time stresses very much as if it were an elastico-viscous material of high viscosity. Assuming that salt flowage is a viscous-like phenomenon we may derive the model ratios for a salt dome.

This is a special case of the general theory like that of the post-Glacial uplift. The movement is assumed to be viscous in character. The forces of inertia are sensibly zero. μ, λ, and τ may be taken as independent model ratios. The derived model ratios of interest then are:

Acceleration:	$\gamma = 0;\ \gamma_g = 1,$
Velocity:	$\eta = \lambda\tau^{-1},$
Force:	$\phi = \mu\gamma_g = \mu,$
Density:	$\delta = \mu\lambda^{-3},$
Stress:	$\sigma = \mu\lambda^{-2} = \delta\lambda,$
Viscosity:	$\psi = \sigma\tau = \delta\lambda\tau.$

The appropriate physical constants of a salt dome in the Gulf Coast region are given by Nettleton [28]:

Depth to mother salt:	20,000 feet
Density of salt:	2.2

Density of sediments:

Depth (feet)	Density
0	1.9
2,000	2.2
4,000	2.27
6,000	2.32
8,000	2.36
10,000	2.39
20,000	2.47

The average area of salt domes in this region is between 1 and 2 square miles.

The age of the mother salt bed is as yet uncertain. Barton [29] presents evidence that it is at least as old as Lower Cretaceous, and possibly older. He also shows that there is evidence that uplift of the domes has been occurring from early Eocene to the present.

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[28] L. L. Nettleton: *Fluid mechanics of salt domes*, Am. Assoc. Petrol. Geol., Bull., vol. 18 (1934) p. 1175-1204.

[29] D. C. Barton: *Mechanics of formation of salt domes with special reference to Gulf Coast salt domes of Texas and Louisiana*, Am. Assoc. Petrol. Geol., Bull., vol. 17 (1933) p. 1025-1083.

According to present radioactive time determinations the beginning of the Cenozoic was about 60,000,000 years ago.

The viscosity of salt is a somewhat uncertain quantity, depending upon whether flowage takes place predominantly by plastic deformation

TABLE 3.—*Properties of original and model salt dome*

	Original	Model Ratios	Model
Radius	10^5 cm.	$\lambda = 5 \times 10^5$	5 cm.
Height	6×10^5 cm.	" "	30 cm.
Thickness of mother salt	To be determined	" "	Assume different values
Time of formation	6×10^7 years or 2×10^{15} sec.	$\tau = 10^{-11}$	2×10^4 sec. (5.5 hours)
Viscosity	Salt: 10^{18} C.G.S.	$\psi = \delta\lambda\tau = 5 \times 10^{-16}$	5×10^2 C.G.S.
	To be determined	$\psi = \delta\lambda\tau$ τ to be measured experimentally	Viscosity of model to be chosen arbitrarily.
Shearing strength	Sediments: 10^8 C.G.S.	$\sigma = \delta\lambda = 5 \times 10^{-5}$	5×10^3 C.G.S.
Density	Salt: 2.2 Sediments: 1.9—2.5	$\delta = 1$	Same as original

within the crystals themselves or by a slow process of solution at points of higher stress and reprecipitations at points where the stress is less. Laboratory experiments are likely to give values due to the former types of deformation, while the infinitely slower natural movement may be to a great extent due to the latter cause.

Here, as in the case of the post-Glacial uplift, we may do either of two things: we may consider the viscosity of salt as known and solve for the proper viscosity for the model, or we may choose a known viscosity for the model and, by determining experimentally the model ratio of time, solve for the viscosity of salt as an unknown. If the laboratory values for salt viscosity are the same as those effective in salt dome flow these two alternative procedures should give the same results.

Gutenberg [30] cites for rock salt viscosity a value of 2×10^{18} C. G. S.

[30] Beno Gutenberg: *Handbuch der Geophysik*, vol. 2, pt. 1 (1931) p. 539. Berlin.

at 18° C. and 2.5×10^{17} C. G. S. at 81° C. Taking an intermediate value, we may consider 10^{18} to be the correct order of magnitude.

There remains one major unknown—the thickness of the mother salt layer. On this there is no direct information. Barton, by an indirect method, arrives at a figure of about 700 feet. This might be taken as

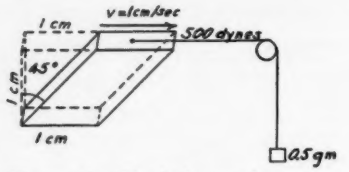

FIGURE 18.—*Physical meaning of a viscosity of* 5×10^2 *C. G. S.*

a trial thickness in the model, but it would be wise to try different experiments in which the salt thickness is varied from one to the other.

We are now able to set up the dimensions for an "original" salt dome and determine the constants for the model. Rather than taking the initial conditions before there was a salt dome, we will define our model ratios in terms of the end-product—the domes as we now see them.

For the conditions given the significant results are that the model equivalent of salt must have viscosity of 5×10^2 C. G. S. The sediments being intruded, if regarded as plastic, must have a shearing strength of about 5×10^3 C. G. S.

To give physical meaning to these figures, imagine in each case a one-centimeter cube of the material, one pair of whose faces are acted upon by a shearing stress couple produced by a weight suspended from a string passing over a pulley. A liquid of viscosity of 5×10^2 C. G. S. means that a shearing stress couple of 500 dynes will produce in the material an amount of shear of unity (shear angle equal to 45 degrees) in one second. The force per square centimeter required to do this is 500 dynes, or approximately the pull of gravity over a 0.5 gram mass.

A strength of 5×10^3 C. G. S. means that plastic flow would occur under a shearing stress equivalent to the pull of a 5-gram weight per square centimeter applied tangentially. If the model is made smaller, or the time shorter, the strength and viscosity of the materials will be correspondingly reduced.

Knowing the physical properties of the model, the experimental procedure would consist in setting up a suitable arrangement of parallel layers, with the equivalent of the mother salt at the bottom. By some manner a doming motion is initiated and allowed to continue.

The more direct experimental procedure is to regard the viscosity of salt and the thickness of the mother layer as unknowns to be determined. The experiment is then set up and varied to determine what model constants are necessary to give the best kinematic representation of a true salt dome. From these models, the length, the time, and the density model ratios are then computed. From the latter the viscosity, *et cetera* of the original may be computed.

The quantities taken here are only approximate and for illustrative purposes. They are, however, the correct order of magnitude. They indicate that of the experiments which have been performed on model representation of salt dome formation, those of Nettleton, employing two viscous liquids, or a liquid and a wax, are the most nearly dimensionally correct.

Impacting Meteorite.—A problem of some contemporary interest is that of the nature of the hole that a large meteorite of given size and velocity would make upon impact with the earth. This is not a problem that lends itself readily to mathematical solution, and the occurrences of such impacts are too few to afford adequate observational data. It remains to be seen if the model technique can be applied. That is, we wish a model of a meteorite to impact a model of the earth's surface in such a manner that the model (including, of course, the hole) is dynamically similar to the original.

This case is the most general possible where inertia, gravity, and resistive forces all play an important part. We choose μ, λ, and τ as fundamental model ratios and obtain the desired ratios:

Acceleration:	$\gamma = \lambda\tau^{-2} = \gamma_g = 1,$
	or $\lambda = \tau^2,$
Velocity:	$\eta = \lambda\tau^{-1} = \lambda^{\frac{1}{2}},$
Density:	$\delta = \mu\lambda^{-3},$
Force:	$\phi = \mu\lambda\tau^{-2} = \mu = \delta\lambda^3,$
Stress:	$\sigma = \mu\lambda^{-1}\tau^{-2} = \delta\lambda,$
Strength:	$\sigma = \delta\lambda,$
Elastic moduli:	$\sigma = \delta\lambda.$

This gives us the necessary model ratios to enable us to set up the model once the properties of the original are assumed. We are here concerned only with large meteorites, so for convenience let us take the original to be a nickle-iron sphere 100 meters in diameter. The velocity of such a meteorite may be expected to be of the order of magnitude of the earth's orbital velocity. At such a velocity a meteorite of this magnitude would possess so great an amount of kinetic energy that it would not be appreciably slowed down by the earth's atmosphere. Meteorites have been observed with velocities as high as 75 kilometers per second. Let us assume a velocity of 50 kilometers per second.

For the model let us use for the meteorite a sphere of lead or of amalgam (to obtain the proper strength) 0.5 centimeters in diameter. This would give for λ the value 5×10^{-5}. The model ratio for velocity then would be

$$\eta = \lambda^{\frac{1}{2}} = \sqrt{5 \times 10^{-5}} = 7 \times 10^{-3}.$$

Taking the velocity v of the original to be 5×10^{6} cm/sec, we may solve for the velocity of the model projectile,

$$v_2 = \eta v_1 = 7 \times 10^{-3} \times 5 \times 10^{6} = 3.5 \times 10^{4} \text{ cm/sec.}$$

This is about 1/3 of the velocity of an army rifle bullet, or somewhere near that of a pistol bullet.

The strength of the model material is $\delta\lambda$ times that of the original. Taking δ to be 1.5 we get

$$\sigma = \delta\lambda = 1.5 \times 5 \times 10^{-5} = 7.5 \times 10^{-5}.$$

If we take the surface materials of the earth to have a shearing strength of 10^{9} dynes/cm^2, the material for the model will have a strength of 7.5×10^{4} dynes/cm^2, and a density of about 3.5.

The strength of the projectile would have to be correspondingly reduced. A nickle-iron meteorite would have a strength of the order of 5×10^{9} dynes/cm^2 and a density of about 8. The model projectile should have a strength of about 4×10^{5} dynes/cm^2 and a density of 12. Some forms of amalgam having these properties might be found. Since the inertial forces are dominant there is little difference at high velocities between the impact of a liquid and of a solid. Hence the effect should be nearly the same if lead were used.

For strict dynamic similarity the modulus of compressibility of the model should be reduced from that of the original by the factor σ. This is probably impossible to do experimentally, but it is doubtful if the error introduced by not doing so is great.

The effect of the air has been neglected. It is known that in the case of the Siberian meteorite the air wave was destructive to local forests. For velocities several times that of sound the air wave is distinctly an after-effect whose magnitude is small as compared with that of the meteorite itself, and may be neglected for present purposes.

For an approximate model of the impact of the meteorite specified we obtain the following: a ball of lead 0.5 centimenter in diameter fired with a velocity of 350 meters per second into the horizontal surface of a material having a density of about 3.5 and a shearing strength of 7.5×10^{4} dynes/cm^2. A material of the latter properties can be made using clay loaded with a proper amount of lead oxide to provide the correct density.

An idea of what might be expected is obtained from the fact that when bird shot from a shot-gun is fired into a bank of soft mud, an individual shot will make a hole about the size of a man's fist. The physical constants in this case are not far from correct for a model of the impact of a large meteorite.

Glacier.—It would be of great importance if a model could be specified and the materials found that would behave in a manner dynamically similar to the movement of ice in glaciers. Insofar as the movement of ice in glaciers is a purely mechanical phenomenon, this can be done and the model ratios for strength, and for viscosity are identically the same as those employed elsewhere.

Strength:	$\sigma = \delta\lambda,$
Viscosity:	$\psi = \delta\lambda\tau.$

Unfortunately, however, the peculiar movement of ice appears to derive its unique characteristics from the unique properties of water. This takes us into the realm of thermodynamics, which, while not outside the domain of model theory or technique, is beyond the scope of the present paper, which is confined to purely mechanical systems. Accordingly, no attempt will be made to specify the properties required for a glacial model. A few remarks regarding the nature of the problem will, however, be made.

The two properties of ice that distinguish it most markedly from ordinary solids are its low melting point and the fact that it expands upon solidification. The internal bodily deformation of glacial ice is not a fluid phenomenon, but rather a plastic flowage involving in part shear displacement within the ice crystals and, in part, the actual melting and freezing by infinitesimal amounts of the ice crystals.

Flow by shearing within the crystals is the familiar plastic deformation of ordinary crystalline solids. The melting and freezing between the crystals is a process peculiar to ice and is the result of both the low melting point and the fact that ice expands on freezing.

Imagine two pieces of ice to be pressed together in a heat-insulated container at 0° C. At the points of greatest pressure the melting point will be depressed below the temperature of the box; melting will occur and water will be produced. The water will flow to the region of reduced pressure where the freezing point is again normal. This may be an infinitesimal distance. The thawing, however, can take place only provided there is a proper supply of heat available—79.7 calories per gram—which is greatly in excess of the heat produced by the mechanical work done on the system. This heat is most readily provided by re-freezing the water produced, thereby obtaining the required 79.7 calories per each gram of water so frozen.

Consequently we have a closed thermodynamic system wherein melting takes place due to the depression of the melting temperature at points of high compressive stress, the water so released flowing to points of lower pressure, and for each gram of ice melted the latent heat being supplied by the re-freezing of one gram of water. The work done, in

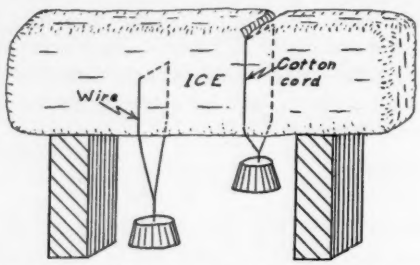

FIGURE 19.—*Block of ice cut by wire and by a cotton cord*

case the pressure is due to weight, is derived from the loss of gravitational potential energy by the system as a whole.

In order for this process to occur it is necessary only that the temperature of the ice be above the depressed melting point. The ice need not be in a heat-insulated container for the process to occur partially, as is shown by a familiar experiment. The experiment consists in taking a large block of ice and bridging it across two end supports. Two metal weights are suspended from the ice, one by a wire passing over the top of the ice and the other by a cotton cord of the same diameter as the wire. The room is at ordinary temperature.

After an hour or so it will be observed that the wire has cut some distance into the ice, but that the water has re-frozen behind the wire leaving no trench. This may be continued until the wire has passed all the way through the ice without cutting it into two pieces. In the case of the cotton cord, the movement is much slower, leaving an open cut behind, for in this case *the water does not re-freeze.*

The explanation is that the metal wire is a good conductor of heat. The ice melts beneath the wire; the water flows around to the top of the wire; the latent heat of melting beneath the wire is supplied by metallic conduction from the freezing water above the wire. This produces a closed thermodynamic system.

Were the wire non-conductive of heat the system could not work. Consequently when the wire is replaced by a cotton cord having heat-

insulating properties, the results are quite different and no re-freezing occurs.

In glacier ice at or near melting temperature this process necessarily takes place. The pressure is supplied by the weight of the ice, each

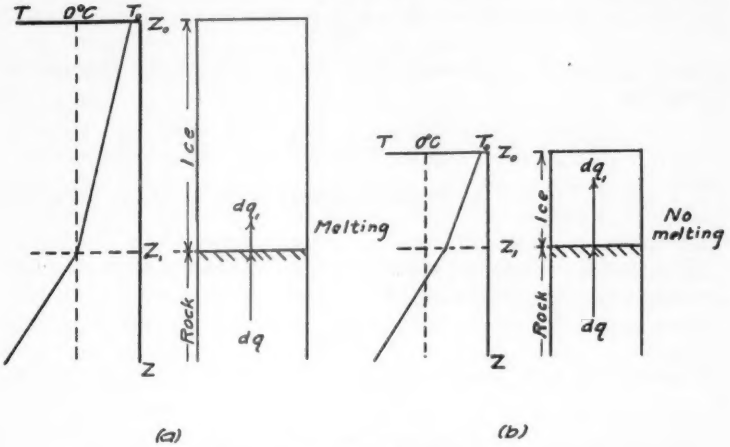

FIGURE 20.—*Temperature gradients*

(a) In rock and in glacier with melting at bottom. (b) Without melting.

infinitesimal amount of melting resulting in a shortening of some length, always in such a direction as to lower the center of mass of the system. The integral of this effect throughout the body of a glacier would of itself constitute glacial flowage.

Another characteristic of glacial motion rendering it unique is the distinct surface of slip between the moving ice and the bounding bed-rock. Glacial motion is essentially laminar, and one usual characteristic of laminar flow is that the velocity is zero at the boundary.

This surface of slip is due largely, if not wholly, to the low melting temperature of ice. There is an important effect, discussed by Lagally,[31] though first called to my attention by T. S. Lovering, due to the flow of heat from the earth's interior.

Consider an ice-sheet of thickness Z_1. Let T_0 be the temperature at the top of the ice and T_1 that at the bottom of the ice. Let dq be the amount of heat coming out of the ground in time dt per unit of area, and dq_1 the amount passing through the ice. Let K_1 be the thermal conduc-

[31] M. Lagally: *Mechanik und Thermodynamik des stationaren Gletschers*, Gerlands Beiträge zur geophysik, supp. vol. 2 (1933) p. 1-90. Leipzig.

tivity of ice and K_2 that of the rock beneath. Let the whole system be in a steady state of flow, that is, not changing with time.

The amount of heat escaping from the ground per unit area in time dt will then be

$$dq = -K_2 \cdot \left(\frac{dT}{dZ}\right)_2 \cdot dt,$$

where Z is positive downward. And the amount passing through the ice will be

$$dq_1 = -K_1 \cdot \left(\frac{dT}{dZ}\right)_1 \cdot dt = -K_1 \cdot \frac{T_1 - T_o}{Z_1} \cdot dt.$$

$\left(\frac{dT}{dZ}\right)_1$ and $\left(\frac{dT}{dZ}\right)_2$ are the thermal gradients in the ice and in the ground respectively.

Now we differentiate two cases: (1) all of the heat flowing out of the ground passes through the ice and

$$dq_1 = dq;$$

(2) part of the heat from the ground does not flow through the ice, and

$$dq_1 < dq.$$

The limiting upper temperature of the ice is 0° C. so that

$$T_o \leqq 0° \text{ C., and } T_1 \leqq 0° \text{ C.;}$$

also

$$T_1 \geqq T_o.$$

In other words the temperature gradient in the ice will always be from a higher temperature at the bottom to a lower one at the top, the limiting case being when both temperatures are the same. The bottom temperature may be equal to or less than 0° C.

Since the amount of heat flowing through the ice is given by

$$dq_1 = -K_1 \cdot \frac{T_1 - T_o}{Z_1} \cdot dt,$$

and since this cannot be greater than the amount of heat coming from the ground, it follows that

$$dq_1 \leqq dq,$$

or that

$$\frac{T_1 - T_o}{Z_1} \leqq \frac{K_2}{K_1} \cdot \left(\frac{dT}{dZ}\right)_2.$$

Since the thermal conductivities are constant and the geothermal gradient is constant by hypothesis, it follows that the thermal gradient

in the ice can never exceed a certain critical quantity. If it is equal to this critical value, the bottom temperature of the ice will be equal to or less than freezing temperature; if it is less than this critical value, the bottom temperature of the ice will be at freezing temperature and the heat from the earth will undergo fractionation, one fraction being conducted by the ice and the other serving as the latent heat of melting for the ice at the surface of contact with the bedrock.

This latter condition is favored by either great ice thickness or high surface temperature. Taking the thermal conductivity of granite to be .008 C.G.S. and of ice to be .006, and the geothermal gradient as 1° C. per 30 meters, we obtain as the maximum value for ice a gradient of 1° C. per 25 meters. This means that for an ice thickness of one kilometer the mean annual temperature would have to be below —40° C. if the bottom of the ice is to be below freezing. For surface temperatures greater than —40° C. the bottom of the ice will be at freezing temperature and a part of the heat from the earth will be spent in producing melting.

This leads us to the conclusion that in most actual glaciers during most of the time the bottom temperature is actually at freezing and that at such times the ice is continuously melting at the contact.

This fact is sufficient to account for the surface of slip at the contact of a glacier with bedrock. In the absence of contrary evidence it appears doubtful that a glacier would slip on the bedrock were the temperature low enough to prevent bottom melting.

It is these unique properties of water in its liquid and solid phases that make it difficult if not impossible to construct a satisfactory model of a glacier.

Owing to the fact that there is a strong similarity between the dissolving and re-precipitation of salts in a saturated solution to relieve local stress, and the melting and re-freezing of water in the manner described, there is a possibility that a suitable glacial model might be devised using some soluble solid and properly controlled interstitial water. Even if possible, the working out of the details presents a research problem of considerable magnitude.

Entire Earth.—This brings us to the last of our models—that of the earth as a whole. As we have remarked before, the strength of the earth is not too certainly known, but the density and rigidity moduli are moderately well known as a function of depth. The over-all modulus of rigidity of the earth as determined in various ways is of the order of twice that of steel. The mean density of the earth is 5.52, ranging from about 2.7 at the surface to about 10 at the center.

The strength of the earth materials increases with increase of pressure and decreases with increase of temperature. Bridgman's recent work

shows that at confining pressures equivalent to a depth of 166 kilometers in the earth, ordinary solids increase in shearing strength commonly by a factor of ten times or more. This is at ordinary temperature, so we do not know how much this would have been reduced at temperatures probably existing at such a depth.

When the shearing stress was equal to the strength of the material, in Bridgman's experiments, any amount of shear could be produced at a stress approximately independent of the rate of shear.

The over-all strength of the earth is probably not greater than that of ordinary steel, so we should not be greatly in error by assuming it to have such a strength.

With this assumption, our problem becomes essentially that of determining how a body having the density distribution and motion of the earth together with the strength of steel should behave under the influence of its own forces, or of such other forces as we might imagine to be impressed upon it. Steel balls of our everyday experience, for example, rebound elastically when dropped upon a steel anvil. Would an elastic earth having the strength of steel rebound if it should collide with a rigid plate at a velocity corresponding to that of the steel ball dropped on the anvil?

To answer these and similar questions we must bring the earth into the domain of our direct observation. We must define the properties of a body having shape, velocity, density, strength, etc., which are strictly similar to the corresponding properties of the real earth, except that the model must be of proper size to be conveniently observed, the criterion of similarity being that of dynamic similarity as previously defined.

For convenience we wish the model to be of a size readily observed—about that of a large terrestrial globe. We wish it to rotate at a rate fast enough to be readily observed, and yet not so fast that one has not time to see the surface details. This would mean a rate of about one or two revolutions per minute.

From such general specifications we determine our fundamental model ratios and derive the remaining ratios that are of interest. Instead of taking μ, λ, and τ to be fundamental, let us choose δ, λ, and γ of density, length, and acceleration. Then we set up the appropriate equations of transformation:

$$\begin{aligned} \lambda &= \lambda \\ \delta &= \mu\lambda^{-3} \\ \gamma &= \lambda\tau^{-2} \end{aligned} \qquad \text{or} \qquad \begin{aligned} \lambda &= \lambda \\ \mu &= \delta\lambda^{3} \\ \tau &= \lambda^{\frac{1}{2}}\gamma^{-\frac{1}{2}} \end{aligned}$$

In this case it should be noted that we imagine our model to be completely removed from the present earth's gravitational field, and to be

influenced only by its own field and that of the corresponding model of the solar system. There will then be a model value of gravity to be determined by the value chosen for γ, which in the general case need not be equal to that on the earth.

Taking δ, λ, and γ as fundamental, the derived model ratios become:

Velocity:	$\eta = \lambda\tau^{-1} = \lambda^{\frac{1}{2}}\gamma^{\frac{1}{2}},$
Angular velocity:	$\frac{\omega_2}{\omega_1} = \tau^{-1} = \lambda^{-\frac{1}{2}}\gamma^{\frac{1}{2}},$
Gravity:	$\gamma_g = \gamma$
Gravitational constant:	$\frac{K_2}{K_1} = \mu^{-1}\lambda^3\tau^{-2} = \delta^{-1}\lambda^{-1}\gamma,$
Force:	$\phi = \mu\lambda\tau^{-2} = \delta\lambda^3\gamma,$
Stress:	$\sigma = \mu\lambda^{-1}\tau^{-2} = \delta\lambda\gamma,$
Strength:	$\sigma = \delta\lambda\gamma.$

For the magnitudes of the fundamental model ratios, the mean radius of the earth is 6.37×10^8 cm. so it is convenient to give the model a radius of 63.7 cm. (about 2 feet) corresponding to a value for λ of 10^{-7}. It also is convenient to let the model have the same density as the original, making δ equal to unity. The proper angular velocity will be obtained for the model if we also choose γ equal to unity; that is, a one-gram mass would weigh the same on the surface of the model as it does on the surface of the earth.

With the fundamental ratios given, the values—

$$\begin{aligned}\delta &= 1,\\ \lambda &= 10^{-7},\\ \gamma &= 1,\end{aligned}$$

we can now write the numerical values of the derived ratios:

Period:	$\tau = \lambda^{\frac{1}{2}}\gamma^{-\frac{1}{2}} = 10^{-3.5} = 3.16 \times 10^{-4},$
Velocity:	$\eta = \lambda^{\frac{1}{2}}\gamma^{\frac{1}{2}} = 10^{-3.5} = 3.16 \times 10^{-4},$
Angular velocity:	$\frac{\omega_2}{\omega_1} = \lambda^{-\frac{1}{2}}\gamma^{\frac{1}{2}} = 10^{3.5} = 3.16 \times 10^{3}$
Gravity:	$\gamma_g = 1,$
Gravitational constant:	$\frac{K_2}{K_1} = \delta^{-1}\lambda^{-1}\gamma = 10^{7},$
Force:	$\phi = \delta\lambda^2\gamma = 10^{-21},$
Stress and strength:	$\sigma = \delta\lambda\gamma = 10^{-7}.$

Then if we know the magnitude for a property of the earth we can determine the appropriate corresponding magnitude for the model. This is done in Table 4.

We now have the specifications for our model of the earth. It is a spheroid of mean radius 63.7 cm. It moves through space under the influence of its own gravitational field and the fields of a model sun, moon, etc., of a model solar system. The mean radius of its orbit is

15.0 km. It rotates once upon its axis in 27.3 seconds and moves in its orbit with a mean translational velocity of 9.44 meters per second, making one complete revolution in 2.77 hours.

The density of the model varies from 2.7 at the surface to about 10 at its center. The over-all shearing strength of its materials is 4×10^2

TABLE 4.—*Properties of the original and a model earth*

Property	Original	Model Ratio	Model
Mean radius	6.37×10^8 cm.	10^{-7}	63.7 cm.
Mean radius of orbit	1.50×10^{13} cm.	10^{-7}	1.50×10^6 cm.
Period of rotation (Sidereal Day)	8.6164×10^4 sec.	3.16×10^{-4}	27.3 sec.
Period of revolution (Sidereal Year)	3.156×10^7 sec.	3.16×10^{-4}	9.98×10^3 sec. (2.77 hours)
Mean orbital velocity	2.98×10^6 cm./sec.	3.16×10^{-4}	9.44×10^2 cm./sec.
Mean density	5.52	1	5.52
Mass	5.98×10^{27} gm.	10^{-21}	5.98×10^6 gm.
Pressure at 100 km. depth	3×10^{10} dynes/cm^2	10^{-7}	3×10^3 dynes/cm^2
Shear strength	4×10^9 dynes/cm^2	10^{-7}	4×10^2 dynes/cm^2
Gravitational constant	$6.67 \times 10^{-8} \frac{\text{cm}^3}{\text{gm. sec}^2}$	10^7	$0.667 \frac{\text{cm}^3}{\text{gm. sec}^2}$

dynes/cm^2. Its gravity will be the same as that of the earth but the gravitational constant of its materials will have to be 10^7 times as large as that for the solar system, or 0.667 cm^3/gm sec^2.

The matter of most concern to us is the strength which has been taken to be the model equivalent of an earth having the strength of cold steel. A material having a shearing strength of 4×10^2 dynes/cm^2 would begin to shear plastically under a shearing stress equivalent to the force due to the weight of a 0.4-gram mass applied tangentially per square centimeter of surface. It is difficult to think of a substance having such a strength but, as a first approximation, very soft mud would be somewhere near correct. That is to say that a sphere of a substance having

a density of 5.52 and the strength of cold steel would, if the sphere were the size of the earth, exhibit properties very similar to those of a sphere of soft mud of radius 63.7 cm.

Perhaps a better way to visualize this is by means of dynamic effects. It was remarked earlier that a sufficiently small sphere of cold steel rebounds elastically when dropped upon a steel anvil. Imagine such a

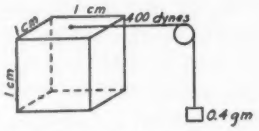

FIGURE 21.—*Physical meaning of a shear stress of 400 dynes/cm²*

sphere the size of the earth to impact against a rigid plane surface while moving at the earth's orbital velocity of 29.8 km/sec. Would it then rebound elastically?

Our model enables us to answer this question. We can imagine the model to impact against such a rigid plane while moving at its orbital velocity of 9.44 meters/sec. From the definition of dynamic similarity the result for the model impact would be similar to what would happen if the earth itself were to have such an accident. A velocity of 9.44 meters/sec. is equivalent to the terminal velocity attained by a body on the earth's surface falling freely from a height of 4.54 meters (14.9 feet).

Now if we can conceive what would happen if a 6.6-ton sphere of soft mud having the density of hematite and a diameter of 4 feet were dropped from a height of 15 feet onto a concrete pavement, we will have a very good idea of what would happen if our earth were to have a similar collision while moving at its orbital velocity of 29.8 km/sec. Instead of rebounding elastically like a small steel ball, it would splatter, although having the strength of cold steel.

SUMMARY AND CONCLUSIONS

At the beginning of this paper we pointed out certain of the difficulties inherent in obtaining a proper idea of geologic structures when these occur on a scale of tens or hundreds of kilometers. Our direct observation of rocks (at any one time) are limited to small specimens possessing great strength, hardness, and rigidity. Yet our geological observations show us that on a large scale these same rocks have been deformed like putty.

Other data confirm the small-scale observations that the magnitudes of the specific physical properties of the earth as a whole are about the same as those obtained by direct measurement on small specimens, embodying great strength and rigidity.

On the other hand, numerous attempts have been made by means of small-scale model experiments to produce results resembling observed geologic structures such as occur in mountain ranges and elsewhere. Invariably it has been found that the experimental results only resembled the original structures in those cases for which the model materials possessed great weakness instead of great strength. This has led to a paradox of an earth whose materials by every known test show great strength, yet whose repeated deformations suggest a material of great weakness.

The resolution of this paradox has been one of the principal purposes of the present paper. To do this, extensive development and use has had to be made of the methods of dimensional analysis. While the present analysis has been confined strictly to mechanical systems, the methods outlined here are equally applicable to thermodynamical, to chemical, and to electro-magnetic systems.

Should this not have been made sufficiently clear heretofore, let it again be emphasized that the methods of analysis employed here are neither original nor new. In his *Two New Sciences* Galileo [32] has an excellent treatise upon the subject of the manner in which the various physical dimensions of a body change with size. That Galileo's understanding of problems such as we have been discussing was by no means superficial can be seen most easily by quoting his own summary.

"From what has already been demonstrated, you can plainly see the impossibility of increasing the size of structures to vast dimensions either in art or in nature; likewise the impossibility of building ships, palaces, or temples of enormous size in such a way that their oars, yards, beams, iron-bolts, and, in short, all their other parts will hold together; nor can nature produce trees of extraordinary size because the branches would break down under their own weight; so also it would be impossible to build up the bony structures of men, horses, or other animals so as to hold together and perform their normal functions if these animals were to be increased enormously in height; for this increase in height can be accomplished only by employing a material which is harder and stronger than usual, or by enlarging the size of the bones, thus changing their shape until the form and appearance of the animals suggest a monstrosity. . . .

". . . Clearly then if one wishes to maintain in a great giant the same proportion of limb as that found in an ordinary man he must either find a harder and stronger material for making the bones, or he must admit a diminution of strength in comparison with men of medium stature; for if his height be increased inordinately he will fall and be crushed under his own weight. Whereas, if the size of a body be diminished, the strength of that body is not diminished in the same proportion;

[32] Galileo Galilei: *Dialogues concerning two new sciences* (1914) p. 130-131. Macmillan Co., New York.

indeed the smaller the body the greater its relative strength. Thus a small dog could probably carry on his back two or three dogs of his own size; but I believe that a horse could not carry even one of his own size."

Sir Isaac Newton [33] contributed further to the subject. Since the time of Sir George Stokes,[34] Helmholtz,[35] and Reynolds,[36] dimensional methods have come into wide use in studies of fluid mechanics. Today the methods are being more and more widely used in all the domains of engineering and of applied physics and chemistry. In the present paper the tools that have been evolving for 300 years are merely borrowed for use in solving problems of a slightly different kind.

By the use of the methods of dimensional analysis the paradox of an earth, apparently both strong and weak, vanishes completely. We see that strength and weakness are purely relative terms devoid of meaning unless the size of the body is specified. Quantitatively, strength is a specific term describing only a specimen of material of a given size; the strength of a body of another size must be determined by analysis. We have made the analysis and learned that quite generally, for a body of material having a given specific strength, the over-all strength of the body taken as a whole decreases with increase of size. Thus small bodies of a given material are strong; large bodies of the same material are weak, and the larger the body the greater its weakness. Conversely, if we are to obtain an idea of the weakness of a large body by observations made on a small one we must artificially weaken the smaller body to bring it into similarity with the larger one. Otherwise we will make the mistake of substituting for an earth whose specific strength is of the order of that of steel our experiences with small steel balls. Steel, as we have seen, when it occurs in the form of small steel balls, is strong; steel, if it occurred in a ball the size of the earth, would be very, very weak.

In the light of what we have now seen it is instructive to appraise some of the better-known model experiments that have been made in attempts to represent mountain-like structures. Most of the experimenters have made the mistake of using materials whose specific strengths were much too great. Bailey Willis,[37] for example, in his discussion of his experiments on Appalachian structures showed plainly

[33] Sir Isaac Newton: *Principia*, Bk. 2, Prop. 32.

[34] Sir George Stokes: *Physical and mathematical papers*, vol. 1 (1880). Cambridge Press.

[35] H. von Helmholtz: *Ueber ein Theorem, geometrisch ähnliche Bewegung flüssiger Körper betreffend, nebst Anwendung auf das Problem, Luftballons zu lenken*, Monatsberichte der Königl. Akademie der Wiss., Berlin (June 1873) p. 501-514.

[36] Osborne Reynolds: *An experimental investigation of the circumstances whether the motion of water will be direct or sinuous, and of the law of resistance in parallel channels*, Royal Soc. London, Philos. Tr. (1883) *or* Sci. Pap., vol. 2, p. 51.

[37] Bailey Willis: *The mechanics of Appalachian structure*, U. S. Geol. Surv., 13th Ann. Rept., pt. 2 (1891-1892) p. 211-289.

that he understood qualitatively the necessity of using materials of less specific strengths than the originals. He failed however to deduce this reduction factor quantitatively, with the result that most of his materials were much too strong. This accounts for the fact that his model structures had to be loaded with an overburden of buckshot equivalent to a depth of rock 1.5 to 2.5 times the length of the model, or if the original were 100 kilometers long the overburden equivalent to that of the model would have been from 150 to 250 kilometers—a geologically impossible amount. Even then the more brittle materials broke into rigid slabs.

Experiments using materials somewhat more nearly correct have recently been made by Kuenen. In some of these experiments Kuenen has employed paraffin floating on warm water, and certain mixtures of paraffin, vaseline, and mineral oil. Nettleton's experiments with salt domes using two viscous liquids of different densities are very nearly dimensionally correct.

The best experiments of all that have come to the author's attention, however, are those of Hans Cloos [38] representing a wide variety of tectonic structures. Cloos actually built his models to conform to simply deduced criteria for similarity. He reasoned that if an original earth feature were composed of rocks strong enough to support a column 10 to 20 kilometers high, a model reduced by a length factor of 1/50,000 would have to be composed of a material that would support a column 1/50,000th as high as that supported by the original materials, or a column 20 to 40 cm. high. It will be noted that this result coincides with the one we have deduced for a similar case, provided the density ratio is equal to unity.

The models made by Cloos display an uncanny resemblance to structures one actually sees in the field. The material used was wet, half-liquid clay too soft to stand without lateral support. In one place Cloos describes it as having the consistency of thick cream. In spite of this it deformed by fracturing into minute blocks so small that the integrated deformation appeared continuous.

The earliest explicit application of the method of dimensional analysis to tectonic structures so far discovered is a paper written in 1912 by Koenigsberger and Morath.[39] Taking their clew from a study by Helmholtz on the theory of hydrodynamic and aerodynamic models, they developed the theory in a simplified form for application to geologic models. They arrived at the conclusion that for a length ratio of 10^{-5}, and a density ratio of unity, the model materials for a mountain struc-

[38] Hans Cloos: *Kunstliche Gebirge*, Nat. u. Mus. Senckenbergische Naturforschende Gesellschaft, Frankfort, pt. 1 (1929) p. 225-243; pt. 2 (1930) p. 258-269.

[39] G. Koenigsberger and O. Morath: *Theoretische grundlagen der experimentallen tektonik*, Zeitschrift der Deutschen Geologischen Gesellschaft, Monat. vol. 65 (1913) p. 65-86.

ture would need to be reduced in strength by a factor of 10^{-5} from the strength of the original.

Koenigsberger and Morath also performed some model experiments, employing for their model materials a mixture of paraffin, vaseline, gutta-percha, Ramsay-fat *(Ramsayfett)*, and machine oil. For loading materials iron filings and lead powder (?) *(Bleipulver)* were used. Iron oxide and chromium oxide were added to give different colors to the separate layers. Experiments were made on scales of 1/100,000, 1/50,000, and 1/25,000.

In an abstract presented to the International Geological Congress in Washington, Koenigsberger [40] again made brief mention of the application of model theory to a study of the size of folds.

The most significant thing about this earlier work is that it seems to have gone very nearly unnoticed. There is evidence that it had the misfortune of being 20 years ahead of its time. The theory developed then was limited in scope but was entirely correct as far as it went.

Bucky's work on mine models, employing a centrifuge to increase the body forces has been mentioned. This is still in a state of evolution, but shows great promise as an experimental technique. It is practically difficult to have a centrifuge that will handle a model more than a foot or so in length, with an acceleration greater than a few thousands of times gravity. The models, however, can be observed by means of a stroboscope, and photo-elastic models can also be used. The method is unique, and may be expected ultimately to yield results of great importance, which are otherwise unobtainable.

Maillet and Blondel [41] have made a more extensive use of this method in working out the properties of a model of the earth. By allowing some unnecessary restrictions and by an unfortunate choice of fundamental ratios they obtained a model of the earth having a radius of 0.04 millimeters which, though dimensionally correct, is much too small to be of use.

There may be others who have contributed to this problem who have been overlooked. If so the oversight is accidental and unintentional. In any case the evidence is that in remote parts of the world the geological professional is already awaking to the importance of so powerful a tool as that afforded by the method of dimensional analysis and correctly made scale models, for the solving of problems that have not yielded satisfactorily to methods of attack previously employed.

[40] G. Koenigsberger: *An experimental tectonic rule proved in the field*, 16th Intern. Geol. Cong. (1936) p. 1000.

[41] R. Maillet and F. Blondel: *Sur la similitude en tectonique*, Soc. géol. France, Bull., sér. 5, t. 4, no. 6-7 (1934) p. 599-602.

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